National Library of Canada
Bibliothèque nationale du Canada

Canadian Theses Service
Services des thèses canadiennes

Ottawa, Canada
K1A 0N4

CANADIAN THESES

THÈSES CANADIENNES

NOTICE

The quality of this microfiche is heavily dependent upon the quality of the original thesis submitted for microfilming. Every effort has been made to ensure the highest quality of reproduction possible.

If pages are missing, contact the university which granted the degree.

Some pages may have indistinct print especially if the original pages were typed with a poor typewriter ribbon or if the university sent us an inferior photocopy.

Previously copyrighted materials (journal articles, published tests, etc.) are not filmed.

Reproduction in full or in part of this film is governed by the Canadian Copyright Act, R.S.C. 1970, c. C-30. Please read the authorization forms which accompany this thesis.

AVIS

La qualité de cette microfiche dépend grandement de la qualité de la thèse soumise au microfilmage. Nous avons tout fait pour assurer une qualité supérieure de reproduction.

S'il manque des pages, veuillez communiquer avec l'université qui a conféré le grade.

La qualité d'impression de certaines pages peut laisser à désirer, surtout si les pages originales ont été dactylographiées à l'aide d'un ruban usé ou si l'université nous a fait parvenir une photocopie de qualité inférieure.

Les documents qui font déjà l'objet d'un droit d'auteur (articles de revue, examens publiés, etc.) ne sont pas microfilmés.

La reproduction, même partielle, de ce microfilm est soumise à la Loi canadienne sur le droit d'auteur, SRC 1970, c. C-30. Veuillez prendre connaissance des formules d'autorisation qui accompagnent cette thèse.

THIS DISSERTATION
HAS BEEN MICROFILMED
EXACTLY AS RECEIVED

LA THÈSE A ÉTÉ
MICROFILMÉE TELLE QUE
NOUS L'AVONS REÇUE

Canada

NL 339 (r. 86/01)

113

0-315-23310-9

National Library of Canada
Bibliothèque nationale du Canada

Canadian Theses Division
Division des thèses canadiennes

Ottawa, Canada
K1A 0N4

PERMISSION TO MICROFILM — AUTORISATION DE MICROFILMER

• Please print or type — Ecrire en lettres moulées ou dactylographier

Full Name of Author — Nom complet de l'auteur

MALCOLM PHILIP BROWN

Date of Birth — Date de naissance
23rd JULY 1962

Country of Birth — Lieu de naissance
GREAT BRITAIN

Permanent Address — Residence fixe
575 MARTON ROAD, MIDDLESBROUGH, CLEVELAND, TS4 3SD
ENGLAND

Title of Thesis — Titre de la these
GLACIOLACUSTRINE SEDIMENTATION IN SOUTH PIPER PASS; NORTHERN ELLESMERE ISLAND

University — Universite
UNIVERSITY OF ALBERTA

Degree for which thesis was presented — Grade pour lequel cette these fut présentée
MASTER OF SCIENCE

Year this degree conferred — Année d'obtention de ce grade
1985 (FALL)

Name of Supervisor — Nom du directeur de thèse
Dr JOHN ENGLAND

Permission is hereby granted to the NATIONAL LIBRARY OF CANADA to microfilm this thesis and to lend or sell copies of the film.

The author reserves other publication rights, and neither the thesis nor extensive extracts from it may be printed or otherwise reproduced without the author's written permission.

L'autorisation est, par la présente, accordée à la BIBLIOTHÈQUE NATIONALE DU CANADA de microfilmer cette thèse et de prêter ou de vendre des exemplaires du film.

L'auteur se réserve les autres droits de publication; ni la thèse ni de longs extraits de celle-ci ne doivent être imprimés ou autrement reproduits sans l'autorisation écrite de l'auteur.

Date
30th September 1985

Signature
Malcolm P. Brown

NL-91 (4/77)

THE UNIVERSITY OF ALBERTA

Glaciolacustrine Sedimentation in South Piper Pass, Northern Ellesmere Island.

by

Malcolm Philip Brown

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH

IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE

OF Master of Science

IN

Geomorphology

Department of Geography

EDMONTON, ALBERTA

Fall 1985

THE UNIVERSITY OF ALBERTA

RELEASE FORM

NAME OF AUTHOR Malcolm Philip Brown

TITLE OF THESIS Glaciolacustrine Sedimentation in South Piper Pass, Northern Ellesmere Island.

DEGREE FOR WHICH THESIS WAS PRESENTED Master of Science

YEAR THIS DEGREE GRANTED Fall 1985

Permission is hereby granted to THE UNIVERSITY OF ALBERTA LIBRARY to reproduce single copies of this thesis and to lend or sell such copies for private, scholarly or scientific research purposes only.

The author reserves other publication rights, and neither the thesis nor extensive extracts from it may be printed or otherwise reproduced without the author's written permission.

(SIGNED) Malcolm P. Brown

PERMANENT ADDRESS:

575 Marton Road,
Middlesbrough, Cleveland,
TS4 3SD, ENGLAND

DATED 30th Sept. 1985

THE UNIVERSITY OF ALBERTA

FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled Glaciolacustrine Sedimentation in South Piper Pass, Northern Ellesmere Island submitted by Malcolm Philip Brown in partial fulfilment of the requirements for the degree of Master of Science in Geomorphology.

.....

Principal Supervisor

Frances J. Hein

.....

External Supervisor

R B Rains

.....

.....

Date 30 Sept 85

Abstract

The United States Range in northeast Ellesmere Island is dissected by a tectonic trench called Piper Pass, which contains three ice-dammed lakes. Abandoned lake sediments on the Hazen Plateau south of Piper Pass, and raised shorelines within the Pass, indicate that more extensive lakes occupied this area in the past.

The research was principally concerned with the stratigraphy and sedimentology of these former lake basins. Related objectives included: 1) Surveying the well defined uppermost shoreline above South Lake (within Piper Pass) to determine whether this shoreline was isostatically tilted and whether it was related to the lacustrine sediments on the Hazen Plateau. 2) Examining the Hazen Plateau sediments and comparing them with a prominent ice-contact delta constructed into South Lake, in order to establish the relationship between sediment deposition and past ice margins within the Plateau Lake basin. 3) Collecting organic remains within these various lacustrine sediments for ^{14}C dating, to determine whether these lakes are either of postglacial or full glacial age.

The three ice-dammed lakes display shorelines up to 30 m above present lake levels. Organics collected from the 30 m shoreline above South Lake were dated at 1585 $\pm$ 190 BP, whereas deposits related to a lake level $\geq$ 16 m dated 495 $\pm$ 170 BP. Collectively, these ^{14}C dates indicate that the changes in lake level occurred in the late Holocene. Aerial photographs taken in 1958 indicate that the lake was 13 m higher than the 1984 elevation. In addition, the absence of observable isostatic tilting along the uppermost shoreline reflects its late Holocene age. The recently expanded South Lake did not extend onto the adjacent Hazen Plateau. Furthermore, there is no evidence of higher lake levels to suggest that South Lake has ever extended out of Piper Pass. Therefore the higher sediments on the Hazen Plateau are not related to a lake within Piper Pass.

Two lacustrine deposits, reaching elevations of 145 m and 156 m above South Lake, mark a lake basin (Plateau Lake) which formerly occupied the northern edge of the Hazen Plateau. Plateau Lake developed when the northward drainage into Piper Pass was blocked by

the Turnabout Glacier debouching from the United States Range. At its maximum, Plateau Lake was 4 km long and 100 m deep. Powerful meltwater streams entered the east end of Plateau Lake from the nearby Divide Glacier and constructed a delta more than 1 km long. Gravels and sands were deposited in proximal parts whereas gently dipping silty-sand foresets prograded across silt-clay bottomset rhythmites in more distal areas. At the west end of Plateau Lake, fine sands, silts and clays were deposited as planar laminations by low velocity currents. Plateau Lake drained in several stages, finally leading to the erosion of sediments deposited at the former ice dam.

Four maximum ^{14}C dates (>33,000 BP, 15,040 ± 625 BP, 20,950 ± 1200 BP and 19,055 ± 2930 BP) obtained on terrestrial organics collected from the two lacustrine deposits indicate that Plateau Lake could be of full glacial age (> 8,000 BP). The ^{14}C dates on these organics suggest that Plateau Lake existed for at least 5,000 years. This is supported by the stratigraphy and sedimentology of the lake deposits, which are used to reconstruct the area's paleogeography.

Acknowledgements

My principal supervisor John England, provided advice and encouragement throughout my masters degree and applied the necessary heavy editing to earlier drafts of this thesis. My external supervisor Fran Hein, reviewed and refined the sedimentology sections. Thanks also to Bruce Rains and Larry Gerard who suggested additional improvements in the final stages.

Financial assistance for the fieldwork was provided by a research grant from the Boreal Institute for Northern Studies, University of Alberta. Logistical support was provided by Polar Continental Shelf Project, Canada Department of Energy, Mines and Resources. I would like to thank my field assistants Aime Stewart and Sydney W.A. Wood for their tremendous patience and hard work under difficult conditions. Dr. J. Wittenberg of the Saskatoon Research Council provided several radiocarbon dates.

Special thanks to Sharon Fowler who typed and retyped the text and made it possible for me to finish within my time limits. Finally, thanks to Emma and Peter who made the long hours in the office during the last two years much more interesting and enjoyable.

Table of Contents

Chapter — Page

1. INTRODUCTION 1

1.1 Previous Research 1

1.2 The Last Glaciation on Northern Ellesmere Island 1

1.3 The Study Area 5

1.3.1 Physiography 5

1.3.2 Geology 7

1.4 Objectives of the Study 10

2. GLACIAL LAKE PROCESSES AND SEDIMENTATION 12

2.1 Previous Research on Lake Sediments 12

2.2 Physical Characteristics of Glacial Lakes 13

2.3 Delta Progradation and Lake Basin Sedimentation 16

2.4 Delta Modification and Resulting Depositional Processes 22

2.5 Conclusions 25

3. RESEARCH METHODS 28

3.1 Field Techniques 28

3.1.1 Southern Piper Pass 28

3.1.2 The Hazen Plateau 32

3.2 Analysis 33

4. SEDIMENT CHARACTERISTICS 37

4.1 Selected Review of Bed Configurations and Stratification 39

4.2 South Lake 41

4.2.1 Ice-contact Delta 41

4.2.2 The Southwest Valley 49

4.2.2.1 Facies I - Ripple cross-laminated fine sands 51

4.2.2.2 Facies II - Planar-laminated sands and silts 55

4.3 Plateau Lake 56

4.3.1 The Christie Section 56

4.3.1.1 Facies III - Cross-stratified sands and fine gravels 58

4.3.1.2 Facies II - Planar-laminated sands and silts 65

4.3.1.3 Facies I - Ripple cross-laminated fine sands 65

4.3.1.4 Facies IV - Rhythmically-bedded silts and clays 69

4.3.2 The Central Valley Section 72

4.3.2.1 Facies I - Ripple cross-laminated fine sands 72

4.3.2.2 Facies II - Planar-laminated sands and silts 76

4.4 Summary 78

5. PALEOGEOGRAPHY AND CHRONOLOGY 79

5.1 South Lake 79

5.1.1 Ice-contact delta 79

5.1.2 The Southwest Valley 85

5.1.3 The Uppermost Shoreline 89

5.1.3.1 Differential Tilting 89

5.1.3.2 History of Shoreline Regression 91

5.1.4 Discussion 91

5.2 Plateau Lake 93

5.2.1 The Christie Section 93

5.2.2 The Central Valley Section 99

5.2.3 Overall Relationships 102

5.3 Chronology 106

5.3.1 Plateau Lake - full glacial age 109

5.3.2 Plateau Lake - postglacial age 110

5.3.3 Summary 113

5.4 Conclusions and Recommendations 113

5.4.1 Conclusions 113

5.4.2 Further Research 115

5.4.2.1 The Study Area 115

5.4.2.2 The Lake Hazen Basin 116

BIBLIOGRAPHY 117

List of Tables

Table		Page
4.1	The Wentworth (1922) Grain Size Scale used for the sediment samples	40
4.2	Estimated mean flow velocities (cm/sec) for flow depths of about 20 cm	42
5.1	Southwest Valley - 6 Sections, 12 Samples	86
5.2	Christie Section - 15 Sections, 32 Samples	96
5.3	Central Valley Section - 10 Sections, 23 Samples	100

List of Figures

Figure — Page

1.1 The study area and key place names on Northern Ellesmere Island and Northwest Greenland. Nares Strait includes Kennedy and Robeson Channels. The Grant Land Mountains include the United States Range. Adapted from England, 1983, p.897 3

1.2 Location map of the study area, showing the southern half of Piper Pass and adjacent Hazen Plateau. 6

1.3 Geology map of the study area, showing the fault zone geological units. Adapted from Miall, 1979, p. ii. 8

2.1 Distribution mechanisms and resulting sediment types in lakes with annual thermal stratification. Adapted from Sturm and Matter, 1978, p. 162. 15

2.2 Classification of ripple-drift cross-lamination based upon the ratio of fallout from suspension to bedload moved by traction. Adapted from Jopling and Walker 1968, p. 982. 18

2.3 Tentative model of a sandy gravelly sediment flow deposit. Adapted from Postma et al., 1983, p. 79. 23

3.1 The Study Area. 29

4.1 The Study Area 38

4.2 Field sketch of the ice-contact delta showing the relative positions of the terraces and meltwater channels. 43

4.3 Field sketch of the southwest valley deposit showing the location of the individual sections. 52

4.4 Field sketch of the Christie section, showing the location of the individual sections. 59

4.5 Field sketch of the Central valley section, showing the location of the individual sections. 73

5.1 Field sketch of the ice-contact delta showing the relative positions of the terraces and meltwater channels. Also shown are previous positions of South Lake and the glacier. 81

5.2 The elevation changes of the uppermost shoreline above South Lake, surveyed in detail from the ice-contact delta into the southeast valley 90

5.3 The approximate maximum positions of Turnabout and Divide glaciers during the existence of Plateau Lake. 107

List of Plates

Plate — Page

4.1 Looking east at the ice-contact delta. Note that the uppermost shorelines are accordant with the highest delta surface (T4) to the lower right of the glacier. 44

4.2 The south meltwater river emerging from beneath T3S(M) and flowing through T2S. The glacier in Plate 4.1 is behind the viewer and South Lake is in the background. 45

4.3 The steeply dipping foresets and overlying topsets of T2N, which is being undercut by the combined North and South meltwater streams. 47

4.4 Section 18 located within T2S. Planar-laminated sands and silts interrupt the foreset beds of pebbles and cobbles in a coarse sand matrix (contact marked by ▽). 48

4.5 Section 20 located 6 m below the T4S surface, composed of fine sands and silts. Note the organics 20 cm below the top of the section which were ^{14}C dated at 1585 ± 190 BP. 50

4.6 The distal part of the southwest valley deposit, with the ice covered South Lake on the extreme right. Note the person on top of the deposit for scale. 53

4.7 Section 6 representing the most distal part of the southwest valley deposit, composed of ripple-laminated fine sands and silts. 54

4.8 Section 4 from the central part of the southwest valley deposit, composed of Facies I sediments interrupted by planar-laminated silts marked by the key. 57

4.9 Looking northwest across the Christie section, the Plateau river and vegetated south side can be seen whereas the top of the main cliff is in shadow. Note the lower western extension of the section in left distance. 60

4.10 Looking west at the main cliff of the Christie section, the top being about 30 m above the modern alluvium from the Divide river in the foreground. From left to right represents proximal to distal. 61

4.11 Section 10, composed of cross-stratified sands and fine gravels, located 4 m below the deposit surface and proximal to the main cliff. 63

4.12 Section 17, composed of cross-stratified sands and fine gravels, located at the top of the main cliff. Organics within fine gravels to the right of the trowel blade were ^{14}C dated at > 33,000 BP. 64

4.13 Section 12, composed of planar-laminated sands, located proximal to the main cliff. This is typical of Facies II. 66

Plate Page

4.14 Section 16, composed of ripple-laminated silty sands, located near the base of the main cliff. Note, organic horizon at the top of the trowel handle which was ^{14}C dated at 15,040 ± 625 BP. 68

4.15 Section 22, located near the base on the south side of the Christie section. The key marks thin organic and yellow fine sand horizons which occasionally occur within the silt-clay rhythmites. 70

4.I6 A close-up of part of section 22. Note the thin organic horizons above the coin and the yellow fine sand horizon near the top. 71

4.17 Looking east at the Central valley section which reaches a height of about 20 m above the stream. Piper Pass is to the right. 74

4.18 Section 6, composed of silts and fine sands, located at the top of the Central valley section. Arrow indicates horizon from which organics were ^{14}C dated at 19,055 ± 2930 BP. 75

4.19 Section 3, composed of planar-laminated silts and sands, located in the middle of the Central valley section. 77

5.1 The ice-contact delta from the air, enlarged from a 1958 aerial photograph. Note the prominent T4N and T3N terraces, however T2 is just forming and T1 is not exposed. Compare with Plate 4.1 and Figure 5.1. 82

5.2 Looking east at the distal end of T1N, which has been cut through and partially collapsed by the river. 84

5.3 Looking east from the top of the Christie section at the Divide river. Note, two terraces which mark former levels of Plateau Lake, and just beyond, irregular mounds which mark a former ice margin of Divide Glacier (black arrows). 94

5.4 Looking down onto the Turnabout junction moraine (m), with the river occupying southeast valley flowing northwards (left) into South Lake. View taken from the top of a ridge above Turnabout valley. 103

5.5 Looking east at the western extension of the Christie section which has sandstone blocks resting on the deposit surface. These are interpreted as ice rafted blocks in the former lake. 105

5.6 Looking southeast into the small gorge on the south side of the Christie section, through which Plateau river is flowing. 112

1. INTRODUCTION

1.1 Previous Research

Environmental research on northern Ellesmere Island began in the early 1950's with glaciological studies on the ice shelves and detached ice islands along the north coast (Koenig et al., 1952; Marshall, 1955; Hattersley-Smith, 1957; Crary, 1958, 1960). Reconnaissance work on the Quaternary geology of northernmost Ellesmere Island by Prest (1952), Hattersley-Smith et al., (1955) and Crary (1956), complemented these initial studies. Subsequently, the International Geophysical Year (1957-58), in turn followed by Operation Hazen/Tanquary, intensified environmental studies, including glaciology and geomorphology, under the direction of the Defence Research Board and the Geological Survey of Canada. During the course of this work, it was proposed that past glaciations in this area were characterized by a southward ice flow from the Grant Land Mountains to the Hazen Plateau (Smith, 1959, 1961; Christie, 1967), and that retreat from the last glaciation had resulted in postglacial emergence of up to 75 m asl (Deane, 1958, 1959; Smith, 1959; Christie, 1967). In addition, Hattersley-Smith (1960a, b, 1961, 1963, 1969 a, b) continued glaciological studies on the contemporary ice cover of northern Ellesmere Island. Finally, during the 1970's there was a change of emphasis towards more detailed paleoenvironmental studies in specific areas (Lyons et al., 1972; Lyons and Mielke, 1973; England, 1974, 1976a, 1978; England et al., 1978, 1981a) which has continued to the present (England, 1982, 1983, Stewart and England, 1983; Bednarski, 1984).

1.2 The Last Glaciation on Northern Ellesmere Island

The maximum ice cover during the last glaciation on northern Ellesmere Island was less extensive than initially thought (England and Bradley, 1978); although some reconstructions still propose an all-pervasive ice cover over the Queen Elizabeth Islands (Hughes et al., 1977; Mayewski et al., 1981). Discontinuous lateral and end moraines form a prominent morpho-stratigraphic boundary along the central and southern parts of the Hazen Plateau

(Figure 1.1). This moraine system, termed the Hazen Moraines, is considered to represent the limit of the last glaciation, approximately 40-60 km beyond the present ice margin (England, 1978). The margin of the northwest Greenland Ice Sheet terminated on the outer coast of Polaris Promontory, leaving an ice-free corridor in northern Nares Strait. Furthermore, the local postglacial isobases over northeast Ellesmere Island and northwest Greenland reflect a strong upward displacement towards the Greenland Ice Sheet. England (1976a) considered this to be consistent with a smaller ice load change on northeast Ellesmere Island (as marked by retreat from the Hazen Moraines), which was isostatically dominated by the considerably larger Greenland Ice Sheet to the southeast.

These field observations from northern Ellesmere Island, plus additional regional data on glacial and raised marine deposits, prompted England (1976b) to suggest a restricted ice cover over the Queen Elizabeth Islands during the last glaciation. This noncontiguous ice cover (the Franklin Ice Complex, England, 1976b) is thought to reflect a period of extreme aridity in the High Arctic, occasioned by the build-up of the Laurentide Ice Sheet to the south (England and Bradley, 1978).

Further research in the ice-free corridor between northeast Ellesmere Island and northwest Greenland indicates the presence of a full glacial sea during the last glaciation (England, 1983). The full glacial sea was apparently in isostatic equilibrium and occupied northern Nares Strait from at least 11,000 to 8,000 BP. Throughout this area it trims weathered till and higher shorelines from earlier glaciations (England et al., 1978, 1981). The pattern of ice retreat on northeast Ellesmere Island is indicated by emergence curves from the full glacial sea, which show that initial unloading (deglaciation) began slowly at 8,000 BP. This was followed by an interval of rapid emergence caused by rapid glacial unloading after 6,200 BP, when a prominent amelioration was in progress (Stewart and England, 1983). In addition to the full glacial sea, earlier workers also suggested the likelihood of terrestrial refugia on northern Ellesmere Island, based on disjunct bryophytes and vascular plants (Brassard, 1971) and entomological data (Leech, 1966).

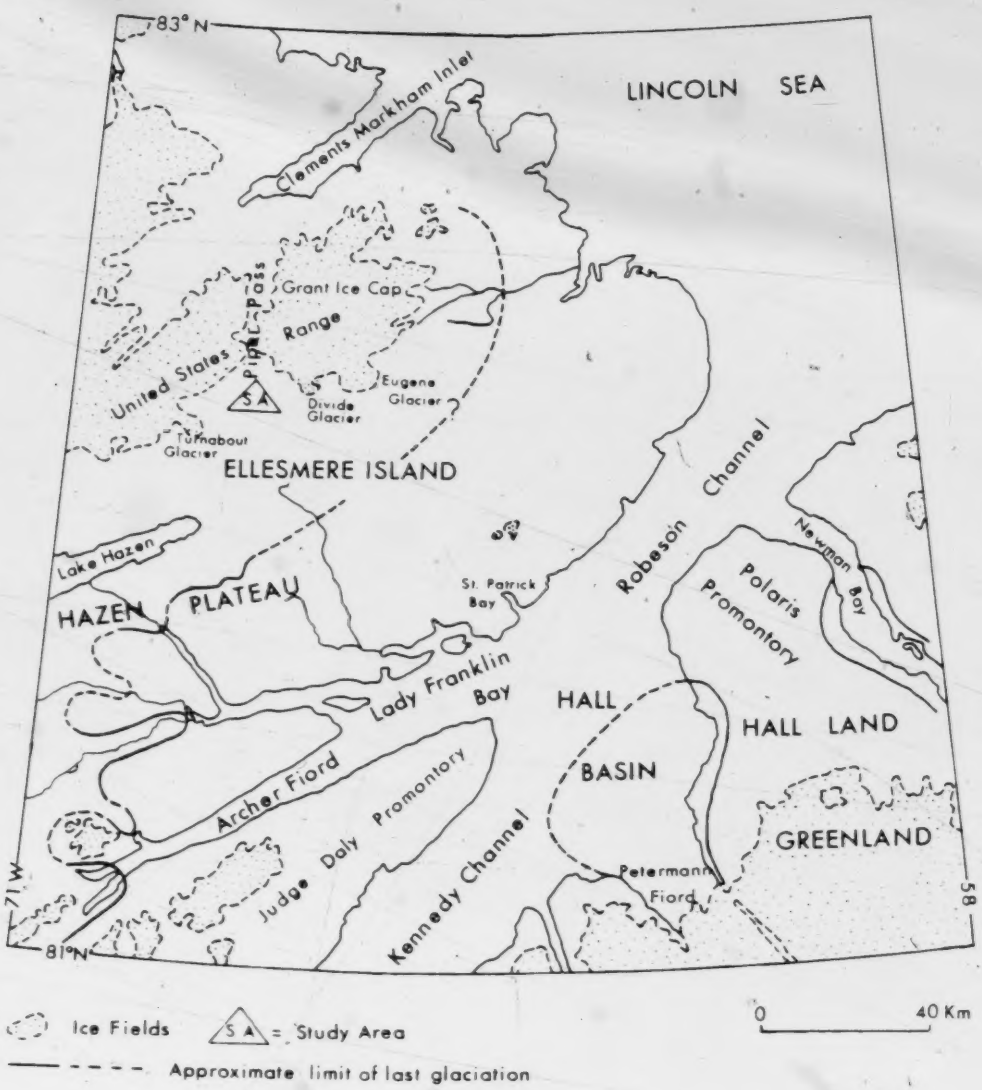

Figure 1.1 The study area and key place names on Northern Ellesmere Island and Northwest Greenland. Nares Strait includes Kennedy and Robeson Channels. The Grant Land Mountains include the United States Range. Adapted from England, 1983, p.897

Recent research in Clements Markham Inlet, north of the Grant Land Mountains, indicates a late glacial history that differs from the Hazen Plateau to the south (Bednarski, 1984; Figure 1.1). The last glaciation in Clements Markham Inlet was characterized by confluent trunk glaciers terminating near the head of the Inlet at about 10,000 BP. Beyond this ice limit there was a full glacial sea that reached 124 m asl. After 10,000 BP, slow retreat was in progress until 8,000 BP, at which time the mouths of confluent valleys at the head of the Inlet were ice-free. After 8,000 BP retreat accelerated greatly, so that the entire lowland became ice-free within 400 years. Relative sea level curves support this pattern of ice retreat, indicating that slow emergence occurred from 10,000 to about 8,000 BP (at a rate of 0.72m/100 years). This period of slow emergence was followed by 'normal' (rapid) postglacial emergence which decelerated to the present (Bednarski, 1984). The pattern of emergence in Clements Markham Inlet is similar to that reported for northeast Ellesmere Island, however the reponses in these two areas are offset by 2,000 years. This chronological difference was likely due to a dissimilar glacioclimatic regime, influenced by the proximity of Clements Markham Inlet to the Arctic Ocean; and the later retreat of both northeast Ellesmere Island and northwest Greenland ice, which lagged behind the initial retreat of ice in Clements Markham Inlet by 2,000 years (England, 1983, in press; England and Bednarski, in press).

On the Hazen Plateau, midway between these two previous field areas, the late glacial history remains undetermined. This part of the plateau is connected to Clements Markham Inlet by a deep trench called Piper Pass (Figure 1.1). At the south end of Piper Pass, on the edge of the Hazen Plateau, former lake sediments are exposed. The purpose of this thesis is to examine these lacustrine deposits in relation to: 1) the contemporary lake system in southern Piper Pass and, 2) the late glacial history of the region. This research therefore connects the recent work in these two areas of northern Ellesmere Island, from which dissimilar late glacial histories have been determined.

1.3 The Study Area

1.3.1 Physiography

Northeast Ellesmere Island can be divided into two physiographic provinces: the Grant Land Mountains to the north and the Hazen Plateau to the south (Figure 1.1). The two provinces are separated by the N.E.-S.W. trending Hazen Fault Zone, which produces a sharp boundary. The study area is located along this fault zone at the northeast end of the Grant Land Mountains. The Grant Land Mountains rise abruptly above the Hazen Plateau to 2,500 m asl. These mountains played a key role in the Quaternary history of the study area because they support expansive, contemporary icefields that attain thicknesses of 900 m (Hattersley-Smith et al., 1969; England et al., 1981b). On the other hand, the Hazen Plateau covers 25,000 km^2 and is largely ice-free. Along its northern margin, in the vicinity of Lake Hazen, it descends to 300 m asl.

The United States Range, a northeast extension of the Grant Land Mountains, is dissected by Piper Pass, a tectonic trench approximately 40 km in length and over 900 m in relief. The Pass is about 3 km in width and largely ice-free at its northern end, however it narrows to about 1 km at its southern end. The central and southern parts of Piper Pass are blocked by two valley glaciers descending from the Grant Ice Cap to the east, and by a third valley glacier descending its western side from the United States Range. These valley glaciers impede the northward drainage from the Hazen Plateau to Clements Markham Inlet, and hence dam three ice-covered, proglacial lakes which fill the southern half of the Pass (Figure 1.2).

The southernmost of the three ice-dammed lakes occurs at 300 m asl and is approximately 8 km long and 1 km wide. It has undergone significant fluctuations in size, as evidenced by a series of raised shorelines and deltas. The largest of the raised deltas was constructed by a small valley glacier in contact with an expanded southern lake. The subsequent reduction in lake size has exposed this ice-contact delta on the east side of the lake, together with extensive lacustrine deposits within both the southeast and southwest valleys draining into

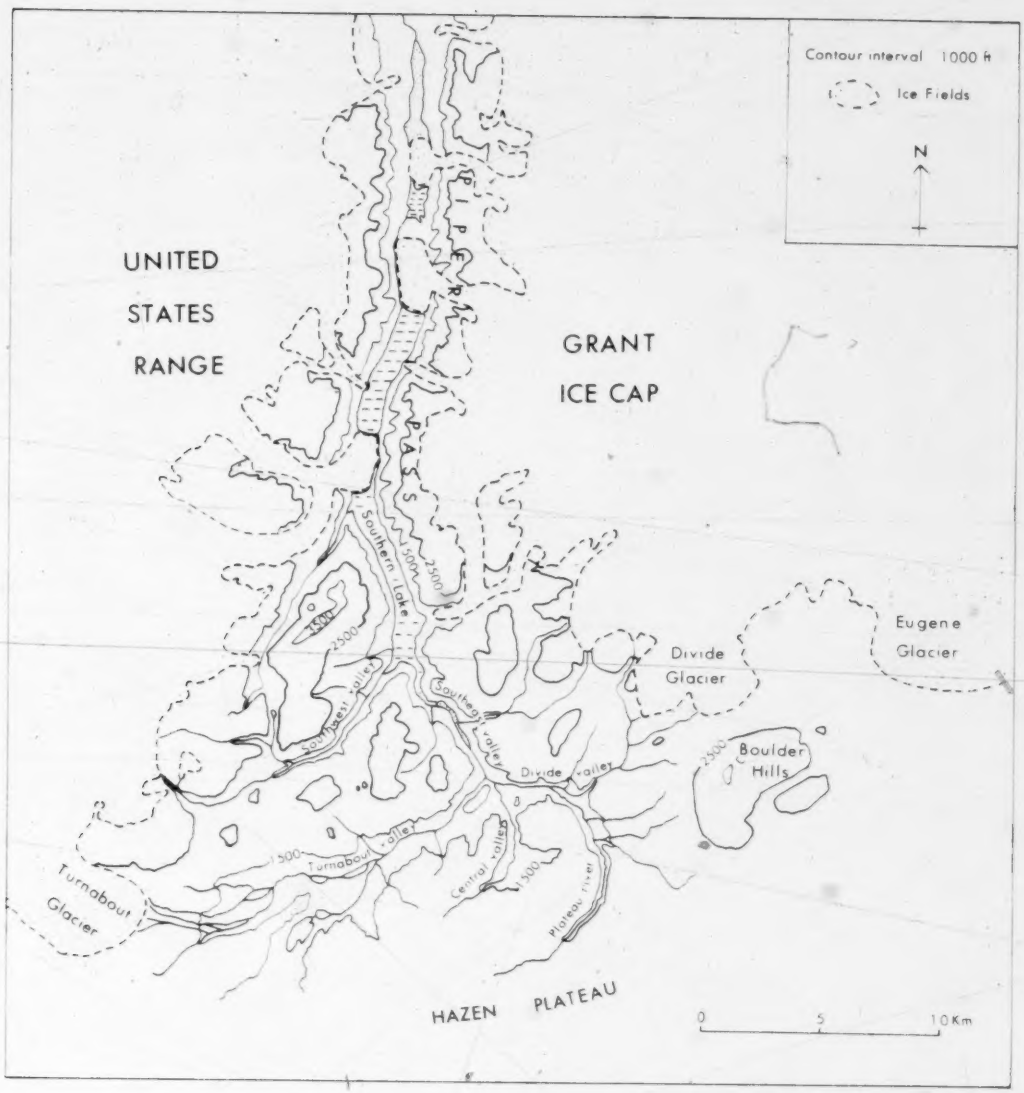

Figure 1.2 Location map of the study area, showing the southern half of Piper Pass and adjacent Hazen Plateau.

it. The uppermost distinct shoreline surrounding the southern lake extends southwards for several kms up the southeast valley towards the Hazen Plateau (Figure 1.2).

Further south, on the northern edge of the Hazen Plateau, isolated lake sediments are exposed and these may also relate to the previously expanded southern lake. Christie (1967, p.21) was the first to describe these sediments:

> Brown sand beds with layers containing twigs, leaves and moss are exposed in banks up to 70 ft high on the edge of the Hazen Plateau upland about 20 miles northeast of Lake Hazen. The plant remains are brown and desiccated, but otherwise appear surprisingly unaltered. The vegetal layers are 2 to 10 mm thick and some parts of the deposit are about one-third organic in volume.

Christie suggested that these detrital beds must have been deposited in a basin formed when the northward drainage in Piper Pass was blocked by ice. Due to the numerous vegetal layers in the lake deposits, he suggested that a climate more favourable to plant growth existed at the time of deposition.

Today the closest glaciers to this former lake are the Turnabout and Divide glaciers (Figure 1.2). The Turnabout is a major trunk glacier which flows southward from the United States Range, terminating along the Hazen Thrust Fault 16 km west of the field area. The Divide Glacier flows southward from the Grant Ice Cap and terminates 5 km east of the field area. These two glaciers provide the majority of meltwater and sediment which flows down the southeast valley and into the southern lake (Figure 1.2).

1.3.2 Geology

The study area lies within the northern Ellesmere Island Fold Belt (Trettin et al., 1972). This region is underlain by lower Paleozoic rocks consisting of sandstones, greywackes, siltstone and shales, which were folded and faulted during the mid Paleozoic, Ellesmerian Orogeny. The lower Paleozoic rocks are superimposed by a succession of upper Paleozoic and Mesozoic strata (sandstones, shale and sandy limestone) which were deformed during the mid-Tertiary Eurekan Orogeny (Trettin, 1971; Figure 1.3).

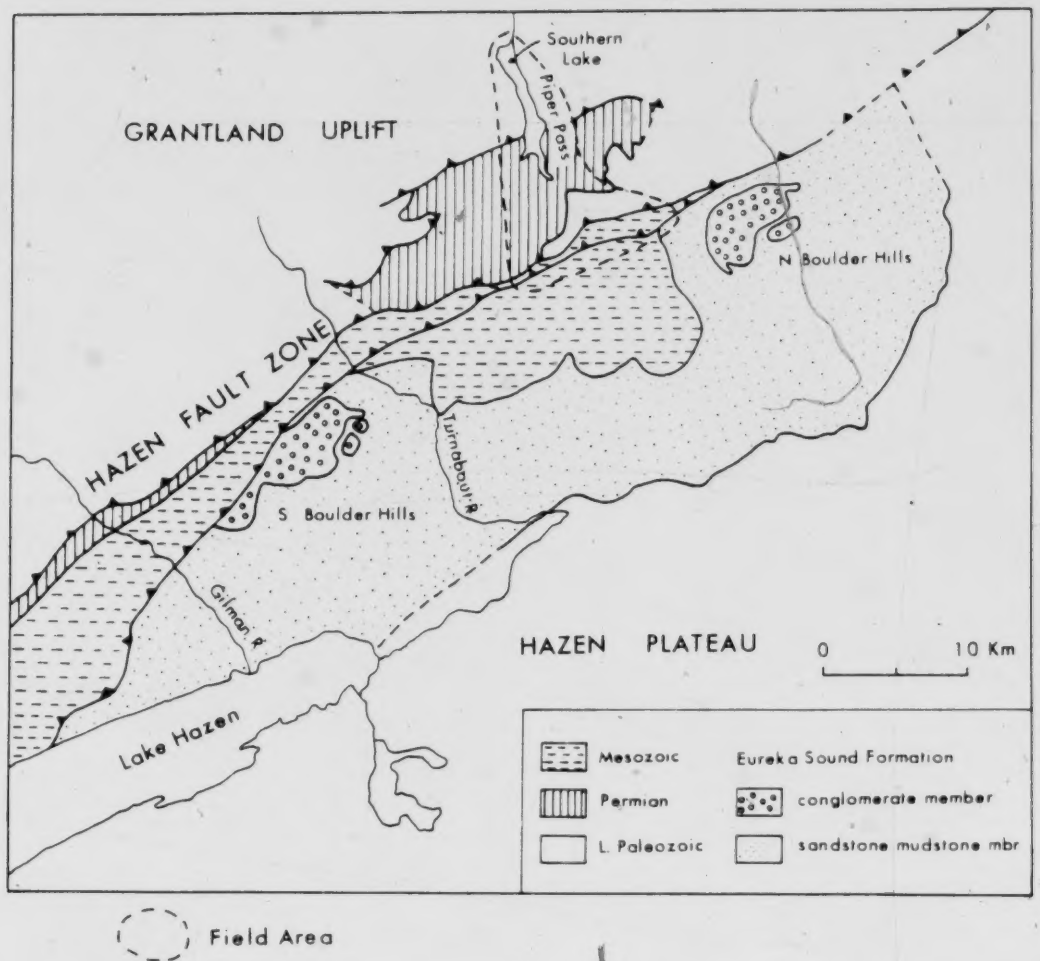

Figure 1.3 Geology map of the study area, showing the fault zone geological units. Adapted from Miall, 1979, p. ii.

The dominant structural feature of the area is the Lake Hazen Fault Zone. The main fault, the Hazen Thrust Fault, is a steep northward-dipping structure which juxtaposes the clastic sediments of the Grant Land Formation (Cambrian and/or Ordovician; Trettin, 1971) against a Permian to Lower Cretaceous sequence of sandstones. A second thrust fault recognized by Christie (1976), which also crosses the field area, occurs between 1 and 8 km south of the Hazen Thrust Fault and is subparallel to it (Figure 1.3). The Permian to lower Cretaceous sequence is, in turn, thrust over Tertiary sediments along this southern thrust fault (Miall, 1979). South of Piper Pass an inlier of Mesozoic strata is present on the south side of the southern thrust fault, and is the only area where Mesozoic and Tertiary strata appear to be in stratigraphic contact (Figure 1.3).

Within the field area Miall (1979) identified two Tertiary units within the Eureka Sound Formation: a lower sandstone-mudstone member, and an upper conglomerate member (Figure 1.3). The sandstone-mudstone member was deposited by small, mixed load streams of the classic meandering type, originating from the Hazen Plateau, south of the basin. The conglomerate member was deposited in response to thrusting and uplift along the Hazen Fault Zone (Eurekan Orogeny). Consequently, there was a shift in provenance and transport direction, with the conglomerates being derived from the Grant Land Mountains north of Lake Hazen. Beds of the sandstone-mudstone member grade up into cobble and boulder conglomerates deposited on an alluvial fan bordering the mountains (Miall, 1979). The conglomerate member is preserved in two locations capping large, isolated hills rising 300 m above the Hazen Plateau. These hills are referred to as the North and South Boulder Hills (Miall, 1979). The North Boulder Hills are a group of irregular and coalescing, dome-shaped hills about 7 km long and 5 km wide, located only 1 km from the present margin of the Divide Glacier (Figure 1.2). The South Boulder Hills are located 20 km southwest of the study area (Figure 1.3).

The geology bordering the southern lake is divided by a northern extension of the Hazen Thrust Fault (Figure 1.3). North of this fault, coarse grained sandstones and blue-grey slates

of the Grant Land Formation weather to produce extensive scree slopes. South of the fault, the massive Permian sandstone (Miall, 1978) weathers into large boulders (some perched on the lake strandlines) and a yellow colluvium. The lake deposits south of Piper Pass are located between the Hazen Thrust Fault and southern thrust fault, and overlie calcareous sandstones and shales of Mesozoic age (Petryk, 1969; Figure 1.3). Although unlithified plant debris of Tertiary age apparently occurs in this area (see Chapter 4), outcrops of it have not been observed.

1.4 Objectives of the Study

The purpose of the research was to provide a better understanding of the late Quaternary environment of the area. Specific objectives included:

1. Surveying and mapping the uppermost shoreline, above the southern lake, which is well defined and extends south towards the Hazen Plateau. This would demonstrate whether this shoreline was related to the lake sediments described by Christie (1967) further south on the Hazen Plateau. Furthermore, if this shoreline is dateable and isostatically tilted, it would add to the present knowledge of glacial unloading in the area. These results could then be compared to the profiles on marine shorelines to the north in Clements Markham Inlet (Bednarski, 1984) and to the south in Archer Fiord/Lady Franklin Bay (England, 1983).
2. Observing the stratigraphy and sedimentology of a prominent ice-contact delta accordant with the uppermost shoreline, constructed into the southern lake. This delta provides an excellent example of ice-contact, glaciolacustrine sedimentation, and it can be compared to the lacustrine sediments south of Piper Pass to determine the nature of sedimentation within that area. As well, the sedimentology of this ice-contact deposit can be compared to research on ice-contact marine sediments conducted at the head of Clements Markham Inlet (Stewart, in preparation).
3. Examining the stratigraphy and sedimentology of the lacustrine sediments south of Piper

Pass. This will establish the environments of deposition during the existence of the former lake and whether they were similar to the ice-contact delta studied within the Pass. The inferred environments of deposition will define the proximity of former ice-margins in this area through time.

4. Investigating the abundant organic remains reported by Christie (1967) within the lake sediments on the Hazen Plateau, and collecting these for ^{14}C dating and faunal identification. These would confirm whether these sediments were either of postglacial or full glacial age - i.e. if the sediments date greater than 8,000 BP they automatically predate any glacial unloading and would therefore be of full glacial age (England, 1983). The fossil mosses from the lake sediments can be compared to the extensive peat depositis found on a nearby fault-block that are presently being studied as a possible terrestrial refugium during the last glaciation (Lafarge-England, in preparation).

Fulfillment of the above objectives should enable the following questions to be answered concerning the late Quaternary paleoenvironment of the study area. Does the chronology of glacial retreat from the last glaciation resemble that of Clements Markham Inlet or northeast Ellesmere Island? Are the lake sediments south of Piper Pass of full glacial or postglacial age, and are they related in any way to the expanded southern lake? This latter question would help define the extent of ice during the last glaciation involving both the Turnabout and Divide glaciers.

2. GLACIAL LAKE PROCESSES AND SEDIMENTATION

The interpretation of glaciolacustrine deposits requires a thorough understanding of sedimentation processes and their relationships. A fundamental characteristic controlling the depositional process is the difference in density between the inflowing meltwater and the glacial lake. The sedimentological relationships of delta progradation and lake basin deposition are governed by the initial interaction, and subsequent circulation, of the meltwater input with the ambient lake water. Additional environmental conditions, such as climate and lake level changes, subsequently determine the degree of post-depositional delta modification, and the resulting lake basin sedimentation.

2.1 Previous Research on Lake Sediments

Research on lake sediments began almost a century ago, although it is only within the last twenty years that a wide range of processes have been presented for the formation of sedimentary sequences. Varved clays were first recognized as lacustrine sediments, and Emerson (1898) was first to describe the deposition of varved sediments from turbid underflows. DeGeer (1912) supported this theory of varve formation in his discussion of varve chronology. However, Johnston (1922) and Antevs (1925) denied that varves were formed mainly from turbid underflows. Antevs assumed that sediment-laden meltwater entering glacial lakes rose to the surface and was distributed by katabatic winds.

Varves were originally thought to form primarily by the settling of silt-sized clasts during the summer melt season, and of predominantly clay-sized particles during winter. Kuenen (1951) later showed that in most freshwater lakes, sediment-laden meltwater can produce and sustain turbidity currents which, in turn, deposit varves characterized by graded bedding. This tended to substantiate the earlier views of Emerson and DeGeer.

The importance of deltaic sedimentation in lakes was first reported in detail by Gilbert (1885) who described the Quaternary deltas of Lake Bonneville (Utah). It has been recognized that the physical characteristics of both the meltwater input and lake margins can vary

considerably. These variations result in a wide range of delta types, although Gilbert-type deltas are well known from modern and Quaternary glaciolacustrine deposits.

2.2 Physical Characteristics of Glacial Lakes

Glacial lakes are lakes which receive significant quantities of glacial meltwater and sediment (Gilbert and Church, 1983). Consequently, they are characterized by relatively high sediment inputs, and high variability of inflow on both diurnal and annual scales (Church and Gilbert, 1975). The sedimentation regime of glacial lakes is influenced by many variables, including: 1) the quantity and size of inflowing sediment; 2) the discharge and density of inflow; 3) the circulation and density structure of the lake water; 4) the basin morphometry; 5) the wind regime and 6) some biological factors (Smith et al., 1982). While some of these variables are site specific, others are controlled by the environment.

Climate directly controls the thermal structure of glacial lakes which, in turn, influences lake circulation. Temperate lakes tend to be dimictic (circulate freely twice a year in spring and autumn), being directly stratified in summer (warmer at the surface) and inversely stratified in winter (colder at the surface) (Hutchinson, 1957; Wetzel, 1975). On the other hand, arctic lakes tend to be either amictic (do not circulate), or cold monomictic (circulate once a year in summer at or below 4°C). Another important aspect of glacial lakes in the high arctic is their permanent or semi-permanent ice-cover, which greatly affects their annual thermal profiles. This ice acts as an insulator during summer by limiting solar heating and wind agitation. As a consequence, lakes tend to remain well below 4°C, maintaining an isothermal character throughout. The lack of wind agitation further reduces lake circulation and the reworking of sediment (Coakley and Rust, 1968; Lemmen, 1984).

Ice-dammed lakes often display distinct characteristics related to the configuration and stability of the ice barrier. They are prone to sudden changes in elevation and extent, leading to the exposure of strandlines recording former lake levels. On northwest Ellesmere Island, all the ice-dammed lakes observed on the ground or from the air showed strandlines well above present

water levels (Hathersley-Smith, 1969a). Recent research in southern Greenland has also demonstrated the presence of well developed raised shorelines marking lake level changes of Neoglacial age. These changes were brought about by the expansion and retreat of a glacier, accompanied by both catastrophic and slow subglacial drainage (Dawson, 1983).

Thorarinsson (1939) suggested that a prerequisite for drainage of ice-dammed lakes is the flotation of the glacier terminus. In this way subglacial drainage is initiated once the water depth at the ice dam reaches nine tenths of the glacier snout thickness, causing the ice barrier to float. Glen (1954) perceived a problem with the barrier flotation hypothesis, because the subglacial outlet may close again once the lake depth fell below nine tenths of the ice thickness. However, under these conditions water existing beneath the ice barrier may continue to drain as it enlarges ice tunnels by thermal erosion. Progressive enlargement of these subglacial tunnels, by the escaping meltwater, may permit lake discharge to increase until the lake has completely emptied (Liestol, 1956). Nonetheless, ice-dammed lakes in polar regions normally drain by flowing over the ice barrier, spilling between the glacier margin and valley slope, or through a threshold situated lower than the ice barrier surface. In some instances (i.e. Axel Heiberg Island) englacial or subglacial drainage has been observed in the High Arctic (Maag, 1969).

Deposition of clastic sediments within a glaciolacustrine environment primarily depends upon the hydrological parameters of meltwater input compared to the lake water properties. Many glacial lakes are temperature stratified during summer, characterized by a warmer surface layer (epilimnion) and a cool, dense lower zone (hypolimnion). The suspended sediment of inflowing meltwater, together with the degree of thermal stratification, will determine the density contrasts and hence circulation and flow patterns within the lake. If the density of inflowing sediment-laden meltwater is greater than that of the lake water, it enters the lake as an underflow or continuous turbidity current. Commonly, during summer, the meltwater is denser than the epilimnion and less dense than the hypolimnion, therefore it will be injected at the epilimnion/hypolimnion boundary as an interflow. If the meltwater is less dense than the lake water, it will flow at the surface as an overflow (Figure 2.1). A fourth type of inflow,

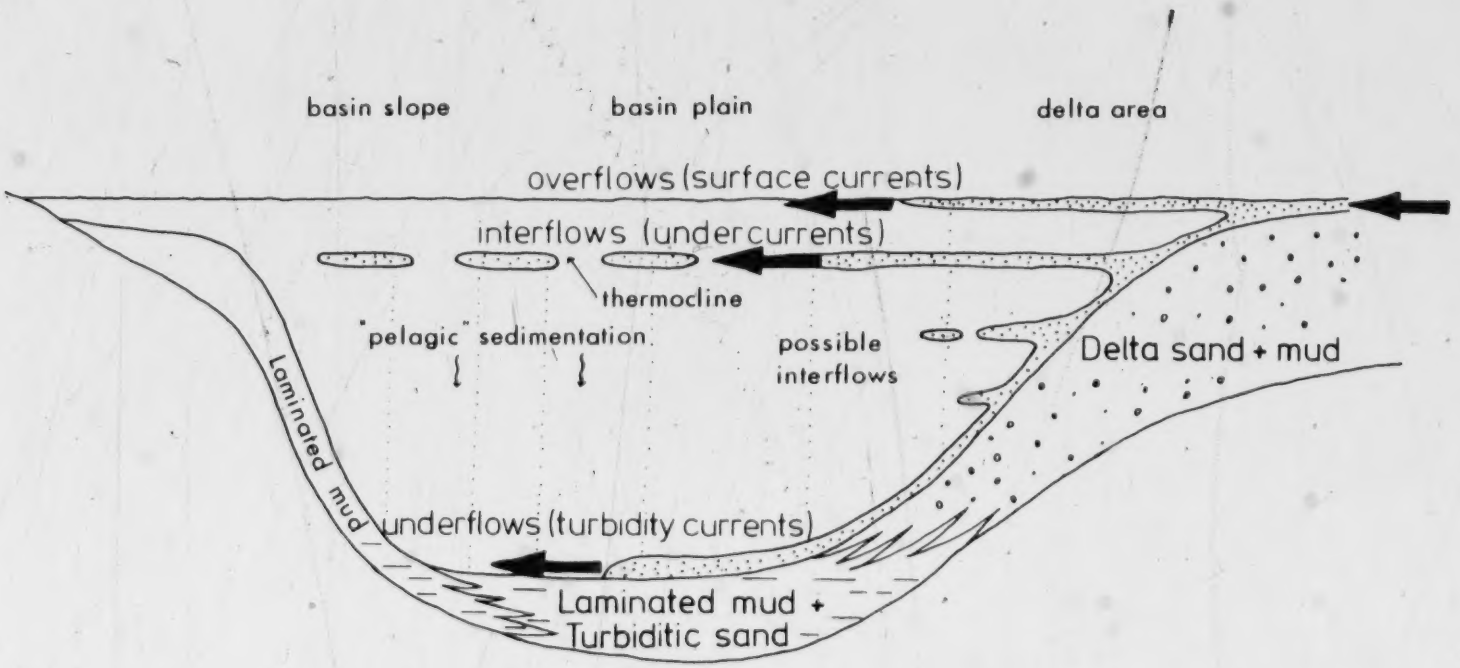

Figure 2.1 Distribution mechanisms and resulting sediment types in lakes with annual thermal stratification. Adapted from Sturm and Matter, 1978, p. 162.

termed homopycnal flow, occurs in weakly stratified or unstratified lakes in which the densities of inflow and lake water are virtually equal. In this case mixing is rapid in all three directions, with inflow quickly losing its uniformity beyond the stream mouth (Bates, 1953).

Sedimentation processes vary significantly with increasing distance from the point of meltwater entry. In proximal environments, sediments coarser than fine sand are deposited rapidly after entering the ambient water, and a delta of predominantly gravels and sands is constructed. Fine sands, on the other hand, are carried into the prodelta environment, whereas silts and clays are transported the farthest, settling-out to form coarse to fine graded beds. Furthermore, fine silt and clay deposition continues during winter throughout the lake, therefore laminated muds are also present within deltas. This simple model of a proximal to distal change from coarse to fine sediments is complicated by many secondary erosional processes on the delta, which can disperse coarse sediments into the lake basin amongst finer-grained sediment.

2.3 Delta Progradation and Lake Basin Sedimentation

Delta progradation occurs at the confluence of rapidly moving river water and quiescent lake water, where a sudden drop in flow velocity and competence takes place. This brings about abrupt deposition of coarse sediment along the delta front, and a slow migration of the upper delta front into the lake (Pharo and Carmack, 1979). Delta progradation results in a typical pattern of horizontal, coarse-grained topset beds, overlying thick, inclined foreset beds (dipping 10-28°) which grade downwards into flat-lying, fine-grained rhythmites (bottomset beds). A zone of channel structures commonly occurs between the well developed foreset beds and the topset beds (Clemmensen and Houmark-Nielsen, 1981).

Foreset beds tend to be composed of several fine-grained sandy, and coarse-grained conglomeratic subunits, often arranged in well-defined sedimentary sequences. The sedimentary structures which make up the cm-to-dm-thick sandy foresets can include planar lamination, ascending and descending megaripple cross-bedding, climbing ripple lamination and isolated

scour-and-fill structures. Descending cross-bedding is formed by megaripples climbing down the delta front, and most often occurs during a simple underflow that is too weak to develop a lee-side vortex. Ascending cross-bedding is thought to form in a lee-side vortex, created by back-flow currents strong enough to transport medium to coarse sand towards and up the delta front in the form of megaripples. During peak discharge and high flow velocities, the continuous avalanche of coarse material down the delta front destroys any megaripples and deposits pebble and cobble-rich foresets (Clemmensen and Houmark-Nielsen, 1981).

In the areas of deltas composed predominantly of sand, foreset beds commonly dip less than 15 degrees and generally consist of ripple-drift cross-lamination (Gustavson et al., 1975). Different types of ripple-drift cross-lamination were described in a kame delta by Jopling and Walker (1968). Each type corresponds to a specific relationship between the bedload transport rate and the rate of deposition from suspension (Figure 2.2). Jopling and Walker (1968) described a number of sedimentary sequences which begin with sinusoidal ripple lamination, and progress upward through type B and type A ripple-drift cross-lamination. This sequence represents an increasing rate of bedload transport relative to the rate of deposition from suspension, and an increase in flow rate. A fourth type of ripple-drift cross-lamination (Type C) accounts for a gradual change in flow and sediment parameters during deposition of the ripple drift. These bedforms serve to point out the temporal and spatial variation in hydrodynamic conditions that control the deposition of glaciolacustrine deltas.

The vertical facies succession at any one point within a glacial lake basin depends upon the sedimentological processes and the rate of ice movement at the time of deposition. Fine-sand, silt and clay transported into a lake basin may be deposited from (a) the waning flow of a turbidity current, or (b) from widespread suspension when turbidity currents are not prevalent. Turbidity currents normally produce graded rhythmites, which are individual units of rhythmic strata with no time or seasonal connotation (Gustavson, 1975). If the silt and clay is deposited as a couplet over one year, then the rhythmite is called a varve.

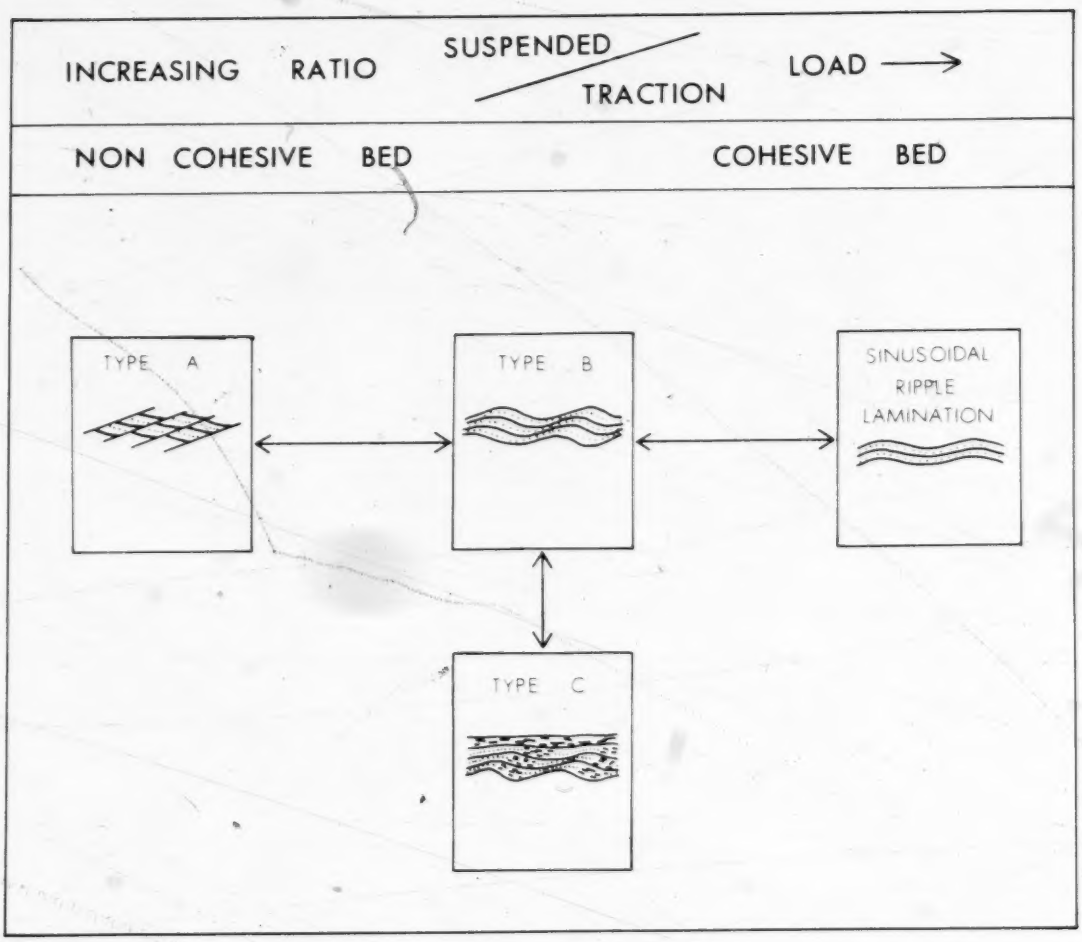

Figure 2.2 Classification of ripple-drift cross-lamination based upon the ratio of fallout from suspension to bedload moved by traction. Adapted from Jopling and Walker 1968, p. 982.

Glacial varves are commonly couplets that can be divided into a summer part, which is normally silt, and a winter part which is clay (Banerjee, 1974). Throughout a glacial lake, varves vary in terms of: couplet thickness, relative thickness of the silt and clay layers, sedimentary structures and grain size (Ashley, 1975). Ashley (1975) divided the varves of Glacial Lake Hitchcock, Massachusetts, into three structural groups:

Group 1 - Clay thickness greater than silt thickness.

Group 2 - Clay thickness approximately equal to silt thickness.

Group 3 - Clay thickness less than silt thickness.

Varves belonging to groups 1 and 2 were formed in still water away from river mouths, where little sediment was received directly from turbidity currents. Group 3 varves were formed relatively near delta fronts, in environments characterized by high sedimentation rates.

Most of the coarse silt layers of varves are deposited by fluctuating turbidity currents and therefore show a variety of current-generated sedimentary structures. Commonly, a single silt layer consists of multiple-graded units each of which show loading structures at its base. These graded units can be explained by the fluctuating velocity of the currents, or by changes in sediment content of an entering stream, possibly caused by the diurnal melt cycle or additional runoff due to storms. The occurrence of ripple cross-lamination appears to be related to grain size, this being a common sedimentary structure in sandy facies, with micro-cross laminations present in coarse silty varves or rhythmites (Banerjee, 1974). There is considerable lateral variation in the thickness of the silt layers, which tend to be thickest close to the delta front and thinnest distal to the meltwater source.

Although clay settles out continuously, it becomes the dominant deposit only during winter when coarse material is less available. The thickness of the clay layer usually remains similar throughout a lake. This spatial difference in the thickness of silt and clay layers, clearly demonstrates the two depositional processes of turbidity currents and vertical settling, which form varves. At any one location within a glacial lake there is often a vertical change in thickness of the varved units. An upwards increase in thickness is characteristic of glacial

sediments during ice advance, whereas an upwards decrease is characteristic of ice retreat (Shaw, 1977).

All glacial lakes do not contain varves or varve-like sediments. The lack of varves has been reported in lakes with a high sediment input and in lakes with a very low sediment input. For example, Gilbert and Shaw (1981) attributed the absence of varves in Sunwapta Lake, Alberta, to the variability of sediment input on a diurnal and subseasonal scale. Intervals of large sediment input dominated sedimentation and therefore masked the annual and semi-annual variability. The few studies carried out on arctic lake sediments indicate that varves are not common. In non-glacial arctic lakes, the absence of varves is generally attributed to very low rates of sedimentation (Coakley and Rust, 1968). Hattersley-Smith (1969a) did report varved clays and silts in three glacial lake deposits on northern Ellesmere Island; however, some studies of contemporary glacial lake sediments failed to recognize varves, although the sediments were clearly laminated (Gilbert and Church, 1983; Lemmen, 1984). It is possible that Hattersley-Smith (1969a) was interpreting temporally variable rhythmites as varves.

There are various opinions as to whether or not thinly laminated deposits can form and survive in periodically draining, ice-dammed lakes. Weidick (1963) stated that sedimentation of "varved character" can only take place in ice-dammed lakes with outflow across stable thresholds. By this Weidick is implying lakes with continuous drainage and not lakes with rapid outbursts. Larsen (1961), on the other hand, noted that in one ice-dammed lake that emptied in southern Greenland, 1 cm varves were disturbed only to a minor degree by water cutting some grooves. Varved clay covering the bottom of an ice-dammed lake that drained catastrophically has also been observed in Iceland (Thorarinsson, 1939).

The finely laminated rhythmites observed in periodically draining ice-dammed lakes are, however, somewhat different from normal varves. This is because the grain-size distribution varies considerably at one site, both in terms of successive sediment layers, and within a deposit of a given season. In addition, these rhythmites are frequently resorted during

and after deposition by wave and current action during outflow of the lake (Maag, 1969).

The deposition of fine sand, silt and clay within a lake basin, determined by the character of meltwater distribution, has been well documented. In contrast, the process of ice-rafting in glacial lakes has received little attention in the literature. Ice-rafting is the deposition of sediment directly from ice into a subaquatic environment, and it is a very important process in ice-contact lakes where icebergs are abundant. The melting of icebergs and lake-ice, deposits coarser sediment into distal lake sediments, and produces several characteristic structures (Thomas and Connell, 1985).

Melting icebergs drop individual clasts from their base, that can range from coarse sand to large boulders. When dirt-laden icebergs breakup and overturn, large quantities of debris are released to the bed, forming lenses or cones of diamicton. Icebergs may become grounded, and subsequently melt *in situ* to produce troughs of diamicton within otherwise fine-grained lacustrine sediment (Thomas and Connell, 1985). In glacial lakes underlain by ice, icebergs can be derived from the lake bottom. For example, Gustavson (1975) noted that in Malaspina Lake, Alaska, some of the emergent icebergs retain on their upper surfaces more than a meter of lake sediment, which is transported and dropped elsewhere in the lake. Consequently, sediment all over the lake bottom receives ice-rafted glacial debris that ranges in size from boulders to clay-sized fragments (Gustavson, 1975).

A second depositional process that is very rarely reported in the literature is aeolian sedimentation. Aeolian sedimentary processes are locally important in arctic regions, especially in outlet valleys from the ice caps where katabatic winds are frequent and intense, and where sandur sediments are exposed (Gilbert, 1982). The amount of aeolian sediment in a glaciolacustrine sequence is extremely difficult to establish. The significance of aeolian sources, however, may be estimated from the accumulation of sediment on ice-covered lakes, although no data exists on its volumetric contribution to glaciolacustrine sediments. In the case of permanently ice-covered High Arctic lakes, the contribution from aeolian sediment is greatly reduced, and probably restricted to the lake margins where a shore lead develops in the

summer.

Recently, sedimentary structures and deposits have been recognized in lake basin environments which cannot be explained by meltwater currents alone. Coarse sediments, and structures indicative of a higher energy environment, have been reported in otherwise fine-grained sediments distal to the meltwater source. Therefore, additional processes, for example winter storms, failure of deltaic deposits and jokulhlaups, are required to account for these sediments (Shaw et al., 1978).

2.4 Delta Modification and Resulting Depositional Processes

A variety of mass movements modify glaciolacustrine deltaic sequences, and transport coarser material into more distal environments of the lake. During peak discharge the upper delta front is the site of heavy sedimentation of coarse sands and gravels, as well as silty sand and organic debris. The slope is maintained in a metastable, oversteepened state, and as more material is added, slumping transports sediment downslope (Pharo and Carmack, 1979). Slumps are initiated by overpressuring and liquefaction associated with rapid sediment accumulation on the delta front. Overpressuring is a build-up of high pore-water pressure, which is usually indicated by abundant water-escape structures in the steeply dipping sediments of the delta slope. Slumping may also result from liquefaction of sediments caused by shocks propagated by the glacier, or calving terminus (Theakstone, 1976).

Following initiation of a slump, the resulting sediment flow may be subjected to several flow stages, especially if the flow accelerates and becomes diluted. Commonly, deposits from sediment (gravity) flows show the effects of plastic flow in the proximal part of a slump, and fluidal flow in the distal part. This produces two main types of sedimentary facies. The first facies (type I, Figure 2.3) represents all the coarse-grained gravelly sediments. These are mainly massive in the proximal location, whereas distally, they become diffusely banded. The second facies (type II, Figure 2.3) represents all the fine-grained, mainly sandy sediments, which fill very shallow troughs. This facies is characterized by normal grading and turbidite-like

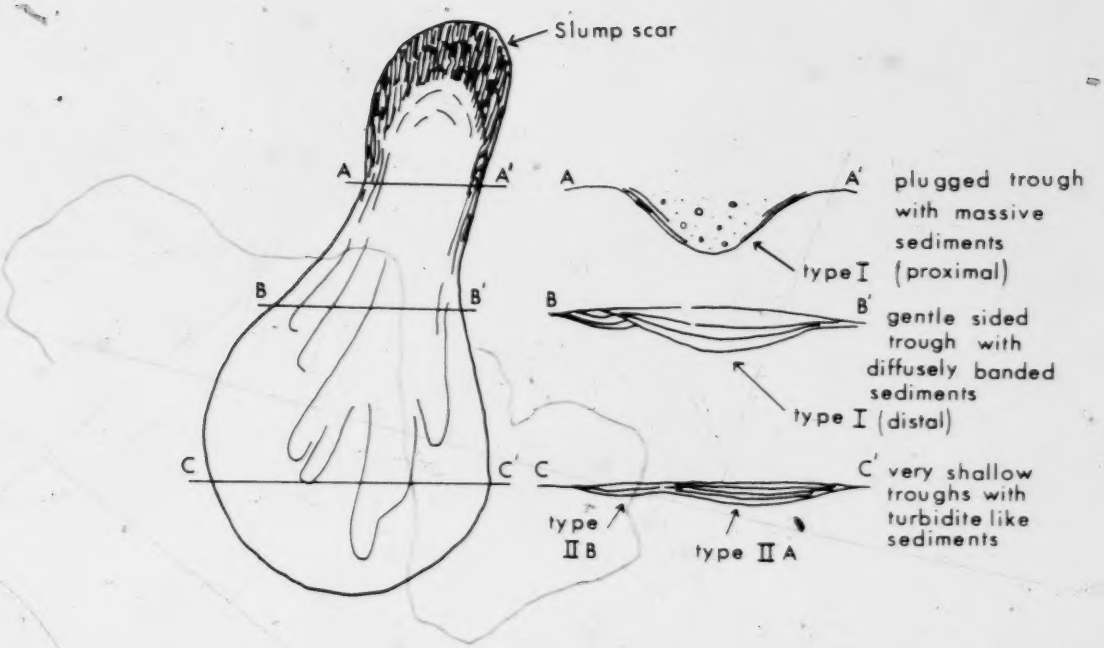

Figure 2.3 Tentative model of a sandy gravelly sediment flow deposit. Adapted from Postma et al., 1983, p. 79.

sequences, which suggest deposition from a turbulent flow. Two subdivisions designate primary and secondary flows, represented by thickly bedded sequences (type IIa) and thinly bedded sequences (type IIb) respectively.

In an ice-contact, deltaic environment, during maximum ablation, high velocity meltwater streams migrate across the delta front and excavate deeply incised channels. Through the combined effects of gravity and of undercutting by the entering stream, accumulated sands and gravels repeatedly experience mass movements which produce elongated lobes of sandy gravel sliding down the delta slope. Some of the sand may become sorted within the sliding mass and travel farther downslope onto the lake bottom as a turbidity current. On the distal part of the delta slope, very fine to medium sand beds are often separated by thin clay layers or deeply-incised erosional surfaces. Sandy gravel intercalations may also be present in this distal area, and they represent the extremities of sediment lobes that slid along the proximal and intermediate parts of the delta slope (Orombelli and Gnaccolini, 1978).

Occasionally, a large mass of material initially deposited on the delta slope becomes unstable and produces sediment density surges. This refers to the episodic formation of strong movements of sediment-laden water that serve to redeposit material from the delta front onto the lake floor. Deposits resulting from such flows show features of turbidites, such as thicker beds of larger mean grain size than those normally occurring at the site, and graded bedding from a sharply defined base. Surge formation may occur during periods of heavy sedimentation in spring and early summer, but also in late autumn and winter when lake levels are sufficiently low to expose most of the delta surface (Pharo and Carmack, 1979).

The above processes clearly indicate that the classical model of a varve, as an annual coarse (silt) and fine (clay) couplet, is not common in glacial lakes. Numerous researchers have presented photographs illustrating coarse-grained deposits within the otherwise fine-grained winter components of glaciolacustrine varves. However, few have realized the significance of these coarse units, and in several cases they have been misinterpreted as summer deposits (Shaw et al., 1978).

Research conducted by Shaw (1977), and Shaw and Archer (1978), in the Okanagan Valley, British Columbia, illustrated sand units within, and in place of, the expected clay components of glaciolacustrine varves. Proximal deposits show rapidly alternating layers of structureless sand, current-bedded sand and laminated silts. The sands represent numerous catastrophic events during the continued sedimentation of silt and clay from suspension. Their deposition towards the end of the melt season or during the winter, suggests that the sands were released by slope failure at the delta front. The sands, alternating with silts, show several properties of grain-flow deposits. These diagnostic properties are the absence or poor development of internal stratification; large transported clasts of silt; and load structures and overfolds in the underlying silts resulting from the rapid loading of unconsolidated (saturated) sediments (Shaw, 1977).

In the distal environment, sand beds also show sedimentary structures characteristic of high sedimentation rates, and are sandwiched between clay layers produced by slow settling during winter. The cross-laminated sands are the product of short-lived, turbidity currents which swept across the glacial lake floor. The resultant turbidity current deposits exhibit the expected characteristics under a waning flow (Shaw and Archer, 1978). The distal turbidity current deposits during winter suggest that regressive slides, which produced grain-flow deposits at more proximal sites, dispersed to form turbidity currents at more distal sites. The sand slides are thought to result from a reduction in internal strength of deltaic sands, due to seepage pressure during the seasonal fall in lake level. Drainage of the saturated sands will increase the stress and may lead to slope failure at the delta front (Shaw, 1977).

2.5 Conclusions

Each sedimentary facies within a glaciolacustrine sequence can be related to changes in sedimentary processes, which are functions of the melt seasons; delta slope instabilities; and distance from the local ice source and/or delta slope. Certainly, the changing distance from an ice front through time will have a pronounced effect on the sedimentary facies. Superimposed

on this general trend are variations in processes and the amount of sediment generated from one year to the next by climatic changes. Although these variations are largely a response to changing weather patterns from year to year,other factors such as changes in drainage catchment and/or internal glacier drainage, and sediment redistribution by slumping, also can cause significant changes in sedimentation.

With a high rate of sedimentation, and slow glacier retreat in contact with a water body, the resultant sedimentary succession is dominated by deltas, some of which may be rotated and lowered during subsequent meltout of supporting ice (England, in press). Down lake from the deltaic sediments, the succession is dominated by proximal rhythmites including sand slide/turbidity current deposits. Where the rates of ice retreat and sedimentation are of intermediate value, proximal rhythmites with sand slide/turbidity current deposits are expected to predominate. Where the retreat rate is high compared to the rate of sedimentation, varved silts and clays are expected to dominate the succession (Shaw, 1977).

In periodically draining, ice-dammed lakes, laterally persistent sedimentary layers are rare. Small-scale crossbedding and graded bedding (normal and reversed) can be found side by side or above one another, although they usually belong to different depositional environments. Lakes in contact with actively calving ice margins, receive large quantities of ice-rafted sediment. Dropstones, dump, and grounding structures, resulting from the melt-out of debris contained in floating and grounded icebergs, commonly interrupt the sedimentary layers. The entire range of glaciolacustrine facies may be present within one sedimentary sequence, which demonstrates the dynamic nature of sedimentary processes operating within a glacial lake undergoing extreme fluctuations.

The processes of sedimentation within a glaciomarine environment have been well documented (Drewry and Cooper, 1981; Domack, 1982, 1983; Wright and Anderson, 1982) and in many respects are similar to glaciolacustrine environments, particularly in the confined setting of glacially-fed fiords (Powell, 1981). It is beyond the scope of this study to compare glaciomarine and glaciolacustrine sedimentation, but it should be noted that research on

glaciomarine sediments at the head of Clements Markham Inlet is also being conducted, and this work will be referred to in a later chapter (Stewart, in preparation).

3. RESEARCH METHODS

3.1 Field Techniques

This thesis is based upon five weeks of fieldwork conducted during the summer of 1984. The study area was divided into two field areas, one within Piper Pass and one south of Piper Pass on the edge of the Hazen Plateau. The first two weeks involved mapping and examining the raised shorelines and deposits above the southernmost lake within Piper Pass, including the southeast and southwest valleys leading to it. This was followed by three weeks of fieldwork in the second area, concentrating on the exposed lacustrine deposits of a former proglacial lake that has entirely drained from the Hazen Plateau.

3.1.1 Southern Piper Pass

An initial reconnaissance established the extent of the uppermost shoreline around the southernmost lake (referred to as South Lake), as well as the extent and exposure of the raised lacustrine sediments (Figure 3.1). The uppermost shoreline around this lake is particularly well preserved and can be traced intermittently for at least 12 km. Its approximate elevation above the lake is 30 m. On the east side of the lake, north of the prominent ice-contact delta, and on the west side of the lake, slopes are steep (50°) and extensively covered by scree, which conceals the uppermost shoreline in many places. Surveying these parts of the uppermost shoreline by a Quickset Level was not possible, both from the view of field safety and the intermittent shoreline preservation. Two field altimeters were used to determine elevation changes along these parts of the uppermost shoreline, taking the present lake as the local base level. Several small deltas and debris cones constructed into the uppermost level of South Lake helped to define the position of this shoreline.

South of the ice-contact delta the uppermost shoreline is continuous (0.5 - 2.0 m wide), extending along the east side of the lake to the mouth of the southeast valley (Figure 3.1). Here, it continues up the southeast valley, although it is more distinct on the west side of

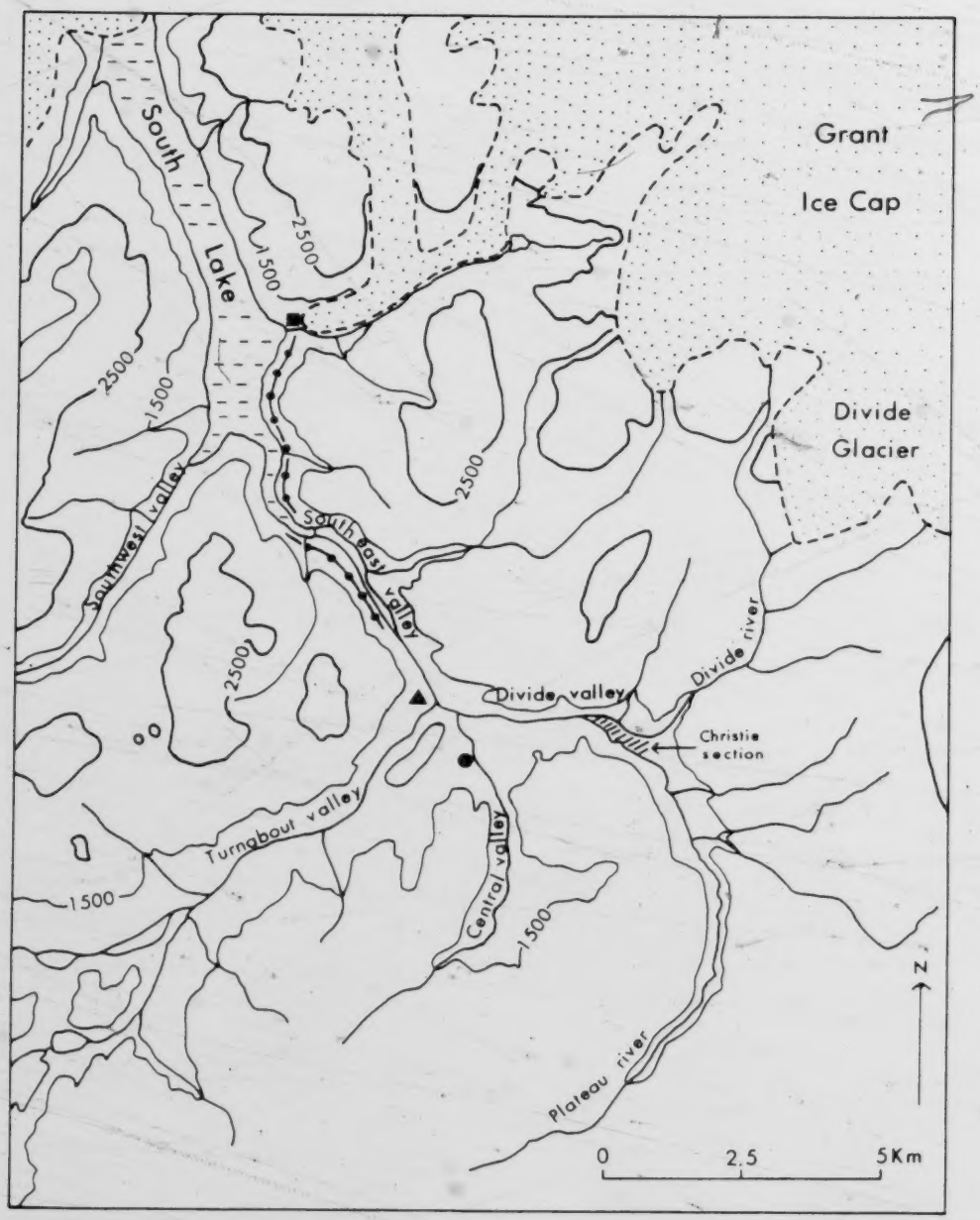

Figure 3.1 The Study Area

● Central valley section ■ Ice contact delta

▲ Turnabout valley moraine Contemporary Ice

—•—•— Uppermost shoreline surveyed in detail Contour interval 1000 ft

the valley, due to the limited scree cover and tributary stream action there. This southern part of the shoreline was surveyed in detail for 6 km. The local base level was taken as the lake edge at the ice-contact delta (base camp 1). A continuous traverse was carried out from the highest terrace on the ice-contact delta (accordant with the uppermost shoreline), up the southeast valley to where the shoreline merges with a former outwash plain (Figure 3.1). If the uppermost shoreline could be dated, then any isostatic tilt along its 6 km length(straight line distance) would assist in refining the pattern of differential postglacial rebound in the area. Altimeter readings extending north of the ice-contact delta to the glacier damming South Lake add to the recorded tilting of the shoreline based on the levelling work. Surveying the uppermost shoreline by level established the accuracy of the altimeter which, for the height of the expanded lake (+30 m), was within ± 2 m of the levelled elevations.

The morphology and sedimentology of the recently abandoned, ice-contact delta was studied in order to characterize the most proximal example of glaciolacustrine sedimentation. The size and construction of the delta was mapped using chain surveying and levelling techniques, and a Brunton compass was used to obtain slope angles of the terrace risers. Primary stratigraphic units were described and photographed, and the sedimentology was based on 20 sections. The sedimentology of individual beds will help to determine the depositional processes that were operating during construction of the delta. Sediment samples were also collected for laboratory analysis to determine the degree of fining within the delta, both laterally and distally from the meltwater source.

Lacustrine sediments were deposited in the southeast and southwest valleys during the uppermost stage of South Lake. As the lake lowered, these sediments were covered by outwash and then dissected and eroded by meltwater streams grading to the present lake. The river occupying the southeast valley receives most of its meltwater from the Divide Glacier. However, it also receives some meltwater from the Turnabout Glacier, plus it drains 350 km^2 of the Hazen Plateau south of Piper Pass (Figure 3.1). Subsequent to the fall in lake level, discharge within the southeast valley, particularly during the 2-3 week runoff peak of late July

and early August, has been sufficient to erode most of the lacustrine sediments. Only minor remnants draped along the valley side exist today, and most of these are covered by coarse outwash which further reduces their exposure.

The southwest valley (Figure 3.1) has received substantially less runoff because it drains an area of only 50 km^2, fed by three small valley glaciers. As a result, raised lacustrine deposits within the valley are better preserved and extend 2.5 km up-valley from South Lake. The stratigraphy and sedimentology of these sediments were described in order to characterize sedimentation in a more distal proglacial environment. This part of the expanded lake does not appear to have been in contact with a former glacier during deposition of the sediments. However, the sediment characteristics suggest that the three glaciers at the head of the valley were more advanced than at present. Six sections between 1 and 1.5 m in height were cut into the exposed sediments, representing proximal and distal sediments from both the top and bottom of the deposit. The sections were cleaned, photographed and described. The descriptions included vertical and horizontal changes in primary sedimentary structures, as well as the texture, thickness and paleocurrents of the beds. Sediment samples were collected for particle size analysis, to help define the primary sedimentary structures and establish facies. The sediments contain abundant organic material and one sample from each section was collected for ^{14}C dating. A top sample from one of the distal sections was dated, and I consider this sample to have been deposited when the lake level was close to the uppermost shoreline.

The same methods of sediment description and sample collection outlined above were used for the lacustrine deposits on the Hazen Plateau. The non ice-contact lake sediments within the southwest valley, and the ice-contact delta along the lake's east shore, can both be compared to the lacustrine deposits on the Hazen Plateau in order to reconstruct the environment of deposition within that former lake.

3.1.2 The Hazen Plateau

Reconnaissance work along the rivers of the Divide and Turnabout glaciers established the location and extent of the lacustrine deposits, as well as related terraces and glacial deposits. The majority of the fieldwork involved examining two lacustrine deposits: 1) the main section originally reported by Christie (1967) at the confluence of the Divide and Plateau rivers; and 2) a smaller deposit within the Central valley (Figure 3.1). The stratigraphy and sedimentology of these deposits were photographed and described from vertical sections, representing proximal to distal sites. One problem related to the sections was establishing whether material was *in situ*, or whether it had been moved downslope by mass movement. For example, the proximal part of the Christie section had been extensively affected by slumping, therefore work concentrated on the steepest part of the deposit. At both of the two principal sites, the accumulation of slope material prevented a basal section from being examined. This was unfortunate, because datable organic material collected from the bottom, as well as the top of the two deposits, would allow the duration of deposition to be established. In addition, the vertical sections cut into the deposits were limited to <1.5 m in height due to the frozen nature of the sediments.

The size and morphology of the deposits were established using aerial photographs, as well as chain surveying and levelling techniques. The elevations of each deposit and their specific sections, were obtained using two field altimeters linked to a local base level at the confluence of the Central and Divide valleys. The elevation of this base level above South Lake had been established previously, therefore all elevations within the study area are reported in metres above lake level.

Minor terraces and a trim line downstream from the Christie section were mapped, whereas a double terrace occurs north of the deposit. The elevations of these features were measured by altimeter to determine how they relate to the prominent lacustrine deposits and whether they may reflect former lake levels. A large moraine at the mouth of Turnabout valley was mapped and photographed, and its elevation determined. This moraine may have been responsible for the blockage of drainage into Piper Pass and the formation of the former

proglacial lake (referred to as 'Plateau Lake') on this part of the Hazen Plateau. Additional mounds of debris between the Plateau and Divide rivers, immediately upstream of the Christie section, were mapped and photographed. These deposits appear to represent a former ice margin of the Divide Glacier (Figure 1.2) which provided the principal meltwater and sediment input into the former lake.

The field observations on these proglacial lake sediments provide the basis for reconstructing the late Quaternary environment of a previously unstudied area connecting Clements Markham Inlet to the outer Hazen Plateau. These data were used to construct an overall model of past glacial and deglacial events in the area. Subsequent quantitative analysis of the sediment samples and surveying work, allows initial in-the-field theories to be supported, refined or rejected. Finally, ^{14}C dating of organic deposits in the sediments provides a chronological framework.

3.2 Analysis

Sediment samples collected in the field were analysed for particle size distribution. Samples chosen for analysis represented *in situ* beds within a sedimentary sequence, the size distribution of which would assist in establishing their mode of deposition. The standard method for particle size analysis of soils was used, although the laboratory procedure adopted varied from this method in several ways (American Society for Testing and Materials (A.S.T.M.), 1982).

An additional pretreatment stage was considered necessary due to the high organic content of many samples. Hydrogen peroxide solution (30%) was added to each sample to remove the organic matter, and the excess evaporated in an oven. Dispersion of each sample (using sodium metaphosphate) was completed using an ultrasonic bath rather than a mechanical stirrer. This method was preferred because it allowed complete dispersion to be visibly determined, and it greatly reduced the possibility of abrasion or fracture of individual particles.

Particle size analysis using a hydrometer records density changes (settling velocities) only within the silt-clay range of a sample. Therefore, prior to hydrometer analysis, each sample was wet-sieved at 4.0 phi (62.5 um) to separate the sand from the silt-clay. Removal of the sand, which is extraneous to the hydrometer analysis, prevents additional interference and improves the accuracy of settling over time, especially for coarse silts which are measured within the first two minutes.

An assumption prior to the hydrometer analysis was that at least 30 g of silt-clay is required to obtain a representative distribution of the fine-grained component. This assumption determined the initial weight of sample used in the particle size analysis. A 50 g sample was used for both coarse sediments (sands) and fine sediments (silty-clays). The coarse samples were unsuitable for hydrometer analysis because of the small silt-clay component. Therefore, 50 g was sufficient to analyse the dominant sand fraction through a nest of sieves, and to obtain a total weight of the silt-clay fraction. The fine samples contained a high proportion of silt-clay, therefore 50 g was sufficient for at least 30 g to remain for hydrometer analysis after the sand had been removed by wet sieving. There is no minimum weight of sand required for dry sieving, therefore the sand components of the fine-grained samples were also analysed through a nest of sieves.

Larger samples, greater than 50 g and up to a maximum of 100 g, were used for silty sands. The maximum of 100 g meant that sediment samples containing less than 30% silt-clay (i.e. less than 30 g out of 100 g) were analysed only by sieving, with a total weight obtained for the fine component. Dry sieving of the sand fraction at 0.5 phi intervals determined both the overall weight of the coarse component, and whether sufficient silt-clay remained for a hydrometer analysis. Some of the samples which weighed less than 100 g were found to contain less than 30 g, but greater than 30% silt-clay, therefore additional sediment (up to the 100 g maximum) was used to obtain the required weight of silt-clay. The pretreatment stages were repeated, and later calculations on these samples were adjusted to account for the additional dispersing agent.

Particle size analysis using a hydrometer is based upon the settling of particles out of solution at velocities determined by Stokes Law. However Stokes Law does not apply to particles less than approximately 1 um (10.0 phi) in diameter. The settling of these medium to fine, clay-sized particles is strongly affected by Brownian motion, resulting in fall velocities slower than those predicted by Stokes Law. Disturbance of the settling column caused by insertion and removal of the hydrometer will enhance this motion, thereby further reducing the measured fall velocities of the clay particles. Therefore, the hydrometer is likely to over-estimate the frequency of the clay component, thereby increasing the calculated sorting of a sample. Fortunately, for most samples, the frequency of clay-sized particles was less than 25%. However, caution is advised when interpreting the upper quartile of silty clays.

Alternative methods of particle size analysis are available, which are less time consuming and avoid the degree of disturbance to the settling column. The most common conventional method is that of pipette analysis, which coincides with the particle size analysis provided by hydrometer (Day, 1965; Mckeague, 1978; Lewis, 1984). More sophisticated and rapid techniques include the Sedigraph 5000D and Coulter Counter (for a comparison of particle size methods see Stein, 1985).

Numerical data from the analyses was plotted on semi-logarithmic paper to show percentages smaller than a given diameter. These cumulative percentage graphs, along with tabulated data, were used to calculate the mean grain size, percentages of sand, silt and clay, and sorting (standard deviation). Collectively, 70 grain size curves were used to compare data from different facies. This particle size data, combined with the primary sedimentary structures, were used to define lithofacies within the deposits and suggest processes responsible for their deposition.

Additional post fieldwork analysis included calculations of the surveying work to determine the profile of the uppermost shoreline, and the heights of other features observed within the study area. Organic samples from the field sites were sieved to remove the mineral matter. Six of these organic samples were selected for ^{14}C dating. The dates provide maximum

estimates on the time of the expanded South Lake, and the duration of sedimentation at the two lacustrine deposits on the Hazen Plateau. The dates are considered maximum estimates because the organic material is composed of allochthonous, terrestrial vegetation (Lafarge-England, personal communication, 1985) whose growth naturally predates its redeposition in the former lake. Finally, three samples were submitted for accelerator dates by Isotrace Inc. (University of Toronto) to determine the range of ages in this allochthonous material. These accelerator dates will be available only after the completion of the thesis.

4. SEDIMENT CHARACTERISTICS

This chapter focuses on glaciolacustrine sedimentation within the two lake basins, and their overall relationships. Sediments deposited in South Lake, Piper Pass, were exposed by a 28 m lowering from the lake's uppermost stage. South of Piper Pass two other glaciolacustrine deposits record a second lake system (Plateau Lake), that formerly occupied both the Divide and Central valleys (Figure 4.1).

Collectively, these glaciolacustrine sediments reflect the former currents and circulation in the lakes, as well as the sediment sizes entering these lakes. Presently, during the summer, meltwater streams transport large amounts of sediment through the study area. This meltwater is more dense than the lake water, therefore the predominant currents transporting sediment into South Lake are unidirectional underflows. It is considered likely that former meltwater streams entering Plateau Lake, and an expanded South Lake, also produced similar underflows.

In a lacustrine environment, unidirectional underflows deposit sediments with several different bedforms which have characteristic stratification. Each bedform is determined by three main variables: 1) sediment size; 2) flow velocity; and 3) flow depth (Harms et al., 1982). Numerous flume experiments have established that bedforms are a lot more sensitive to changes in flow velocity than flow depth. In addition, flow depth causes substantial changes in bedform only when it is close to the bedform height. In natural flows, where greater flow depths occur, the main affect of increasing flow depth is the formation of larger bedforms (Harms et al., 1982). Flow depth in a lake refers to the thickness or height of an entering current. This is difficult to estimate from the stratification and sediment size because the flow depths are likely to be greater than bedform heights, and therefore have a negligible effect on the bedforms. For these reasons the stratification and sediment size can be used to reconstruct relative flow velocities in a lake, however, flow depth can be considered to be constant. The three main variables stated above change progressively through a lake, therefore stratification types can be discussed as a continuum from one end member to the other (Harms et al., 1982).

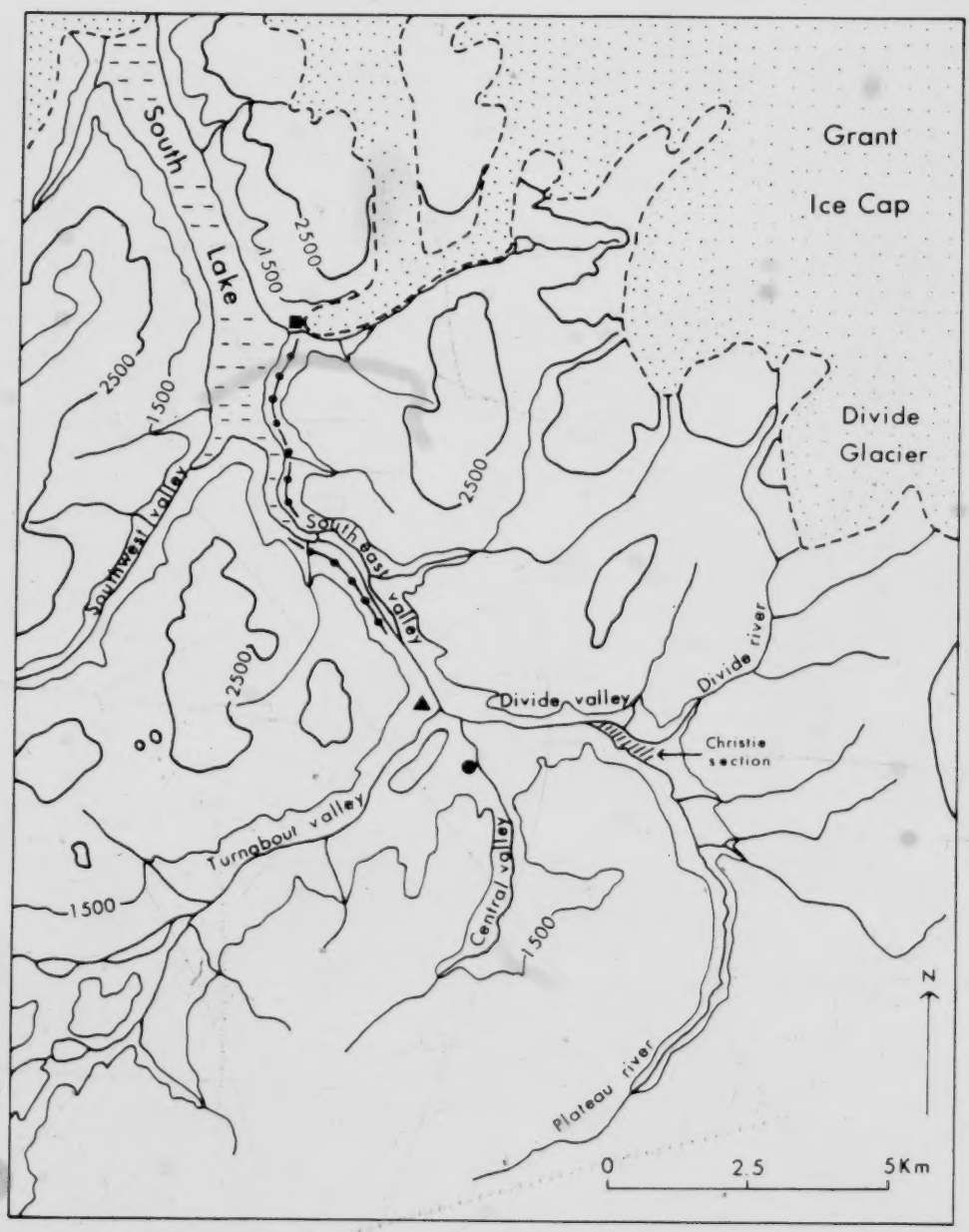

Figure 4.1 The Study Area

● Central valley section ■ Ice contact delta

▲ Turnabout valley moraine Contemporary Ice

—•—•— Uppermost shoreline surveyed in detail Contour interval 1000 ft

4.1 Selected Review of Bed Configurations and Stratification

In areas of lakes affected only by slow-moving currents (less than about 15 cm/sec), sediments smaller than fine sand are deposited from suspension without subsequent traction. This produces plane beds composed of very thin (<0.5 cm) laminations. This planar lamination may grade downward into congruent lamination, which develops from silt and clay draping over a nonplanar (rippled) surface (Harms et al., 1982).

As current velocities increase, the ratio of traction load to suspended load also increases. This produces a continuum of stratification, beginning with sinusoidal ripple-lamination, proceeding through type B ripple-drift cross-lamination, and ending with type A ripple-drift cross-lamination (Jopling and Walker, 1968). Sinusoidal ripple-lamination is the product of current velocities slightly higher than the threshold for bed traction (ca. 20-25 cm/sec). This is sufficient to form an undulating surface, but complete small ripples do not develop. Type B ripple-drift cross-lamination is indicative of increased traction, sufficient to produce small ripples. Deposition from suspension remains high enough for subsequent ripples to climb at a large angle, thereby preserving the complete small ripples. When current velocities are high enough to cause greater bed traction than deposition from suspension, the stoss side of small ripples is partially eroded by the passage of the next ripple. This produces type A, ripple-drift cross-lamination, which is the preserved fraction of ripple foresets. This sequence of stratification can occur in a wide range of sediment sizes, from fine silt up to coarse sand (Table 4.1).

At higher velocities (>40 cm/sec), in sediment of medium sand up to gravel, small ripples are replaced by sand waves (two dimensional large ripples, Harms et al., 1982) and dunes (three dimensional large ripples, Harms et al., 1982). The size of these bedforms ranges upward from spacings of about 50 cm and heights of several centimeters. With increasing bedform development over time, there is a gradation from sand waves (straight crests) to dunes (irregular crests). This transition brings about a change from large-scale tabular, to large-scale trough, cross-stratification. In stronger currents(>70-120 cm/sec), upper-stage plane beds

TABLE 4.1

The Wentworth (1922) Grain Size Scale used for the sediment samples.

U.S. Standard Sieve Mesh #		Millimeters	Microns	Phi ø	Wentworth Size Class
		4096		-12	
		1024		-10	Boulder (-8 to -12ø)
use		256		-8	Cobble (-6 to -8ø)
wire		64		-6	
squares		16		-4	Pebble (-2 to -6ø)
5		4		-2	
6		3.36		-1.75	Granule
7		2.83		-1.5	
8		2.38		-1.25	
10		2.00		-1.0	
12		1.68		-0.75	
14		1.41		-0.5	Very coarse sand
16		1.19		-0.25	
18		1.00		0.0	
20		0.84		0.25	
25		0.71		0.5	Coarse sand
30		0.59		0.75	
35	1/2	0.50	500	1.0	
40		0.42	420	1.25	
45		0.35	350	1.5	Medium sand
50		0.30	300	1.75	
60	1/4	0.25	250	2.0	
70		0.210	210	2.25	
80		0.177	177	2.5	Fine sand
100		0.149	149	2.75	
120	1/8	0.125	125	3.0	
140		0.105	105	3.25	
170		0.088	88	3.5	Very fine sand
200		0.074	74	3.75	
230	1/16	0.0625	62.5	4.0	
270		0.053	53	4.25	
325		0.044	44	4.5	Coarse silt
		0.037	37	4.75	
	1/32	0.031	31	5.0	
Analyzed	1/64	0.0156	15.6	6.0	Medium silt
	1/128	0.0078	7.8	7.0	Fine silt
by	1/256	0.0039	3.9	8.0	Very fine silt
		0.0020	2.0	9.0	
Pipette		0.00098	0.98	10.0	Clay
		0.00049	0.49	11.0	
		0.00024	0.24	12.0	
or		0.00012	0.12	13.0	
		0.00006	0.06	14.0	
Hydrometer					

replace both small ripples, and sand waves/dunes as the stable bedform. Upper-stage plane beds develop in all sizes of loose sediment, when the velocity is sufficient to transport all the sediment as traction load, and thus prevent any ripple development. A second type of planar-bedding is lower-stage plane beds. These are the stable bed configuration under a narrow range of low current velocities, in sediments too coarse (> 0.5 mm) for small ripple development (Harms et al., 1982).

The stratification and grain size of glaciolacustrine sediments reflect the processes responsible for their deposition. These characteristics can be compared to phase diagrams, which are compiled from flume experiments, and used to estimate the velocities of former meltwater currents (Table 4.2). Therefore, the glaciolacustrine deposits within the study area indicate the nature of the former lake environments and their relationships to past ice margins. Most importantly, the ice-contact and proglacial sediments of South Lake, provide a useful comparison to the sediments of Plateau Lake. For a complete review of bed configurations and resulting stratification, see Harms et al., (1982), and Middleton and Southard (1984).

4.2 South Lake

4.2.1 Ice-contact Delta

Retreat of the valley glacier on the east side of South Lake led to the construction of a prominent ice-contact delta (Plate 4.1). The delta consists of four terrace levels, which correspond to four stages of lake lowering. The highest terrace (T4) is 29 m above present lake level and is accordant with the uppermost shoreline. Subsequent stages of the lake are indicated by lower terraces at 21 m (T3), 16 m(T2) and 10 m (T1) (Figure 4.2). The delta is ice-cored throughout, and as a result, the terraces are undergoing rapid modification due to thermal erosion by two lateral meltwater streams (Plate 4.2).

The primary stratigraphic units within the delta terraces are subhorizontal, coarse-grained, topset beds, and inclined foreset beds (dipping between 22° and 29°). Finer

TABLE 4.2

Estimated mean flow velocities (cm/sec) for flow depths of about 20 cm

Stratification Type (Facies)	Sediment Size in mm (phi) <0.10 (<3.5)	0.10-0.14 (3.0)	0.50-0.64 (1.0)	1.30-1.80 (-0.5)
Planar lamination (II_2 and IV)	<15	<20		
Sinusoidal ripple lamination (I_1)	15-30	20-30	20-25	
Type B, ripple-drift cross-lamination (I_3)	35-50	40-45	25-30	
Type A, ripple-drift cross-lamination (I_4)	60-70	50-60	30-35	
Large-scale cross-stratification (III)			40-120	50-100
Upper-stage plane beds (II_1)	80-100	70-120	120-130	>200

Sources: Harms et al. (1982), Middleton and Southard (1984).

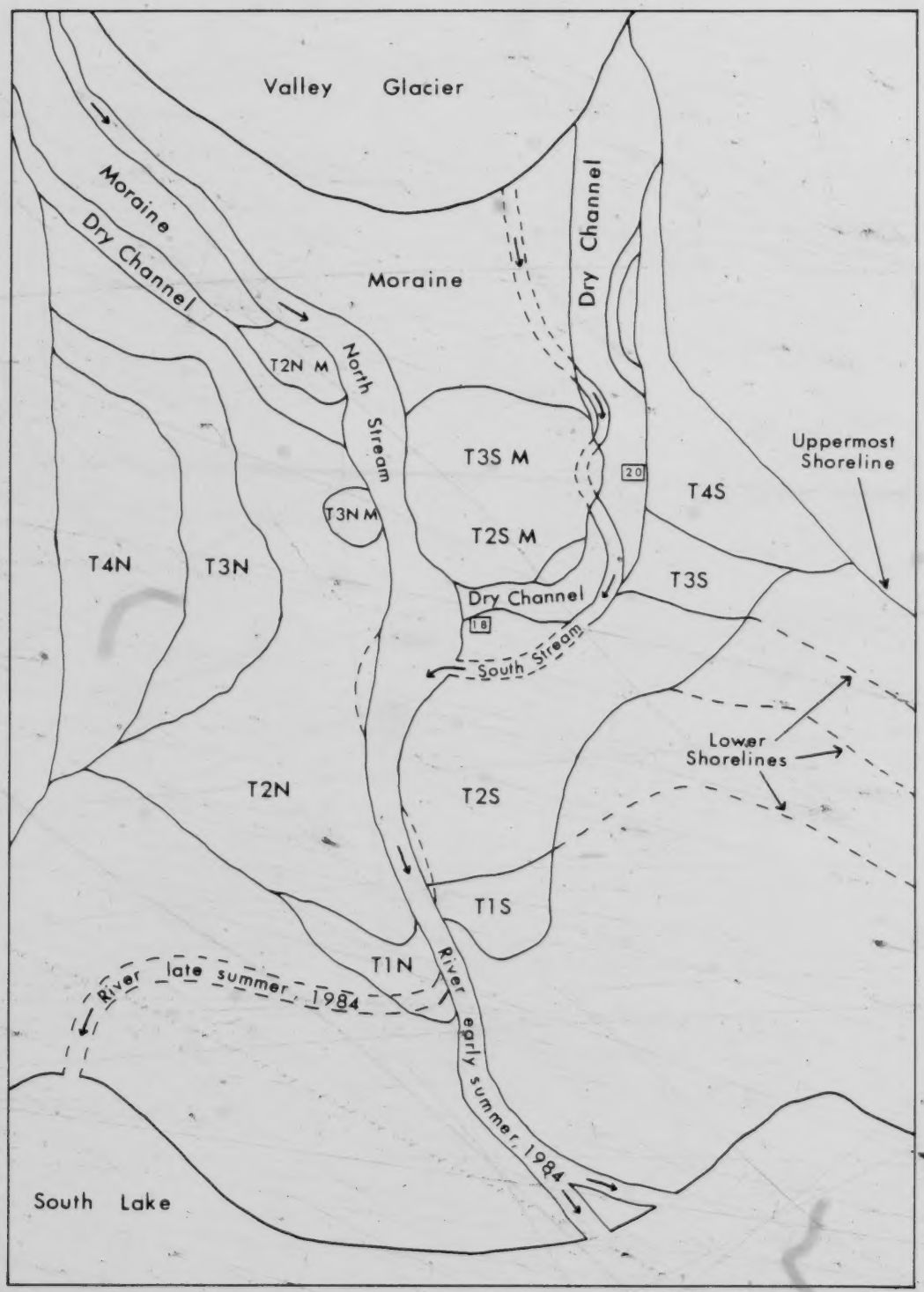

Figure 4.2 Field sketch of the ice-contact delta showing the relative positions of the terraces and meltwater channels. [20] individual sections referred to in the text.

Plate 4.1 Looking east at the ice-contact delta. Note that the uppermost shorelines are accordant with the highest delta surface (T4) to the lower right of the glacier.

Plate 4.2 The south meltwater river emerging from beneath T3S(M) and flowing through T2S. The glacier in Plate 4.1 is behind the viewer and South Lake is in the background.

grained, flat-lying, bottomset beds were not observed, whereas stagnant ice left by the glacier underlies the foreset beds (Plate 4.3). Because the topset and foreset beds are similar throughout the delta, different facies could not be identified, and therefore the stratigraphy and sedimentology are described for the delta as a whole.

The foreset beds tend to be composed of alternating units of pebbles/cobbles, and coarse sands/fine gravels. The foresets are unsorted and lack recognizable structures which, together with the grain size, suggests that they were deposited by high velocity underflows (>150 cm/sec). These underflows were powerful enough to bring about the continuous avalanche of coarse sediment down the delta front and preclude any megaripple development (Clemmensen and Houmark-Nielsen, 1981).

Finer-grained foreset beds occur occasionally within this depositional sequence (Plate 4.4). These are composed of alternating units of medium sand, and fine sand/coarse silt, which contain rare gravel lenses. These foresets consist of inclined plane beds (23° to 28°), varying considerably in thickness (0.5 cm to 10 cm). They are considered to represent upper-stage plane beds deposited from traction by turbidity small too powerful (>70 cm/sec) for current ripple development. A return to the deposition of pebble and cobble-rich foresets can produce loading structures in the finer sediment, whereby grey sandy silts are squeezed up into the yellow sands.

The topset beds are composed of flat-lying pebbles, cobbles and boulders in a fine gravel matrix. The topsets contain channel structures, and the terrace surfaces aremarked with paleochannels. This indicates deposition by high velocity meltwater distributaries on a former sandur. Rare lenses of fine sand draped between the clasts probably resulted from channel avulsion, followed by settling in small pools on the terrace surface.

There is a progressive fining of sediments, both laterally and distally, from the delta apex. For example, within the foreset beds, this change is marked by fewer cobbles and boulders in the T1 terraces, however, the matrix (coarse sand/fine gravel) remains similar throughout the delta. In areas peripheral to the delta apex, sequences of silty sand can occur. Section 20, 6 m below the T4S surface (Figure 4.2), is 54 cm high and composed of brown silts

Plate 4.3 The steeply dipping foresets and overlying topsets of T2N, which is being undercut by the combined North and South meltwater streams.

Plate 4.4 Section 18 located within T2S. Planar-laminated sands and silts interrupt the foreset beds of pebbles and cobbles in a coarse sand matrix (contact marked by ◸).

and yellow fine sands (Plate 4.5). The lower 24 cm (unit 1) consists of sinusoidal, ripple-laminated fine sand, which is transitional to type B, ripple-drift cross-lamination at the base. The ripple-lamination grades upwards into 10 cm of planar-laminated, sandy silts (unit 2). The top 20 cm (unit 3) is composed of fine sand and organics, however the bedding is indistinct due to loading structures. Unit 3 is interrupted by a 3 cm unit of undulating planar-laminated silts (Unit 4; Plate 4.5).

The above sequence is indicative of rapid deposition from suspension by low velocity turbidity currents (<40 cm/sec). Initially, the currents were strong enough for bed traction, however, the absence of type A, ripple-drift cross-laminations indicates that deposition from suspension remained dominant (unit 1). A decrease in velocity below the bed traction threshold (20 cm/sec), produced a gradation to planar-laminated silt (unit 2). The turbidity currents' competence increased, and brought a return to fine sands containing organics (unit 3). It is not possible to identify the lamination type in unit 3 because of the convoluted bedding. These convolutions were probably caused by loading from gravels and pebbles deposited on top of the saturated sands. The deposition of unit 3, however, was interrupted by calmer conditions, during which the 3 cm silt unit was deposited (unit 4). Organics collected from 20 cm below the top of section 20 were ^{14}C dated at 1585 ± 190 BP (S-2637).

The ice-contact delta serves as a good example of coarse, clastic sediments deposited by retreating ice, with the foresets and topsets overlying buried ice. The delta was initially constructed at the highest lake stage because T4 is equivalent to the highest shoreline (28 m). The subsequent fall in lake level produced a staircase of four delta terraces, and finally led to deep incision and thermal erosion throughout the delta.

4.2.2 The Southwest Valley

During the highest stage of South Lake, additional sediments were deposited into a prominent valley that leads to its southwest shore. This is referred to as the southwest valley (Figure 4.1), and three small glaciers occupy this watershed 11 km to the southwest. The

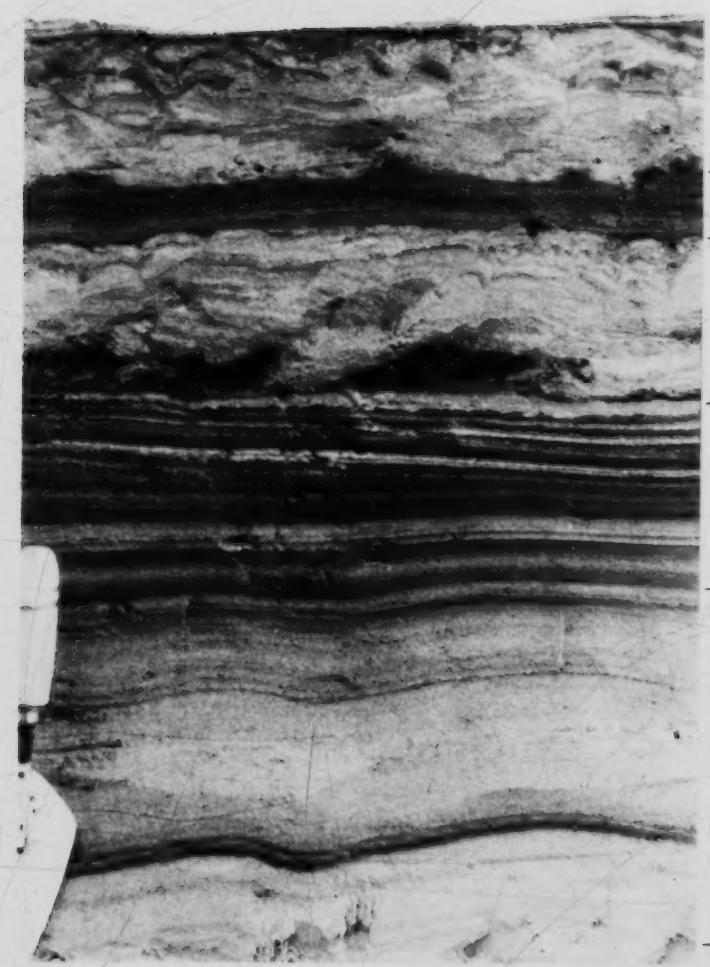

Plate 4.5 Section 20 located 6 m below the T4S surface, composed of fine sands and silts. Note the organics 20 cm below the top of the section which were ^{14}C dated at 1585 ± 190 BP.

sediments in the lower part of southwest valley reach a thickness of 16 m above South Lake and are predominantly gently dipping, fine sands, containing redeposited terrestrial organics (Plate 4.6). Examination of six sections through these sediments indicates the presence of two facies: 1) widespread, ripple cross-laminated, fine sands interspersed with; 2) planar-laminated sand and silts in lesser amounts.

4.2.2.1 Facies I - Ripple cross-laminated fine sands

Facies I consists of moderately sorted, fine sands and organics, which are the products of migrating small ripples. A variety of ripple cross-laminated stratification was recognized in Facies I, and grouped into three subfacies: 1) type A, ripple-drift cross-lamination; 2) type B, ripple-drift cross-lamination and 3) sinusoidal ripple-lamination. These subfacies represent variations in ripple geometry, which developed in response to changing flow conditions. Section 6 shows Facies I in which the three subfacies can be recognized (Plate 4.7).

Section 6 is 70 cm high and represents the distal part of the lacustrine deposits in southwest valley (Figure 4.3). The lowermost 22 cm consists of fine to very-fine sand, in the form of type A, ripple-drift cross-lamination (subfacies 1). This stratification indicates that current velocities were high enough (50-60 cm/sec) to transport the sand from the stoss to the lee side of a ripple at a greater rate than the addition of new sand to the bed. Therefore, bed traction dominated, and the ripples migrated at a small angle of climb. The fine sands are the sigmoidal foreset laminae remaining after partial erosion of the stoss side of the ripples.

An oxidized, erosional contact separates subfacies 1 from a 20 cm unit of sinusoidal, ripple-laminated silty sands (subfacies 3). There is a gradation within this 20 cm unit, from stratification transitional between subfacies 2 and 3 at the base, to stratification transitional between subfacies 3 and plane beds at the top (Plate 4.7). The sinusoidal ripple-lamination indicates that current velocities were only slightly above the threshold for bed traction, and that deposition from suspension was the dominant mode of

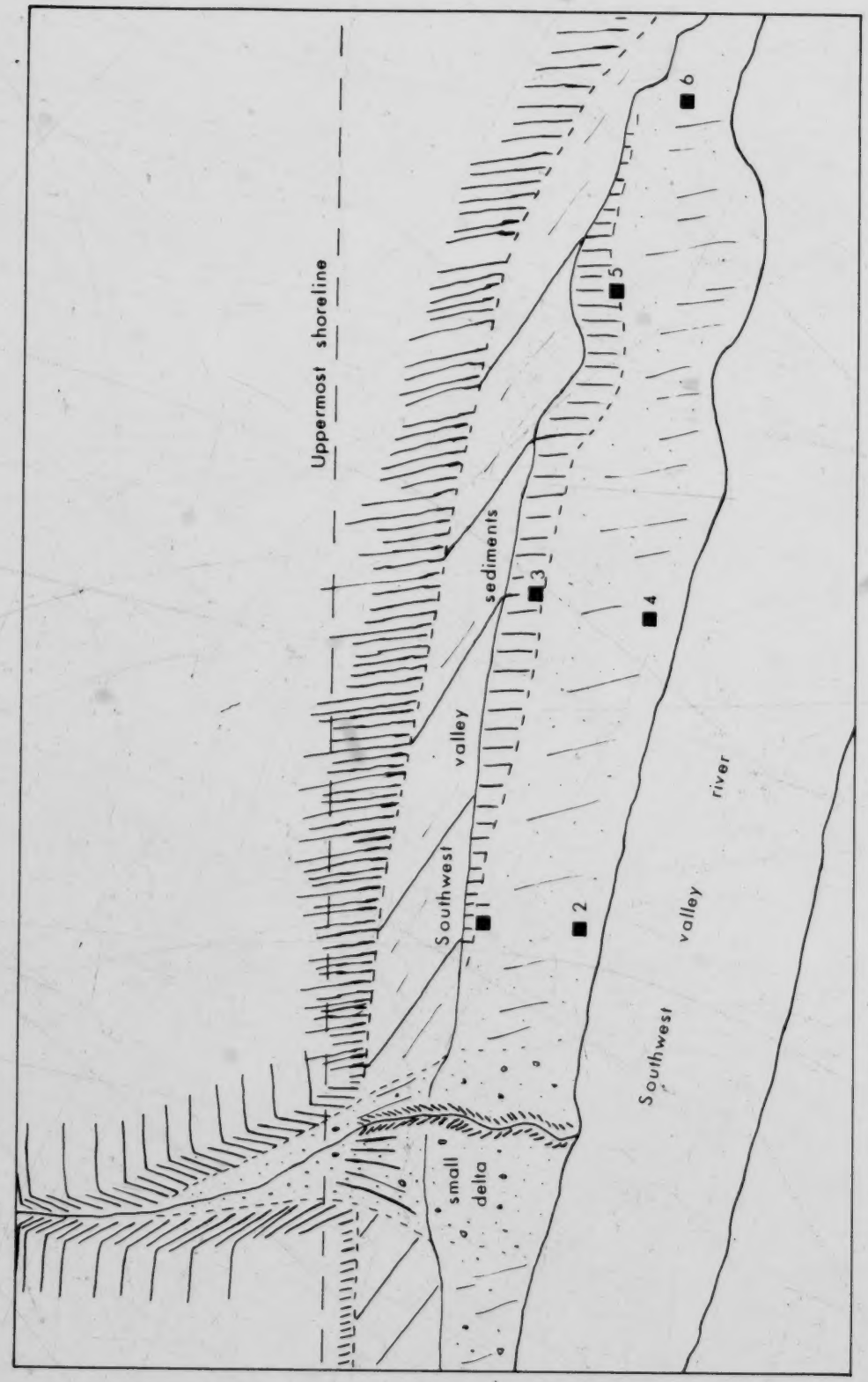

Figure 4.3 Field sketch of the southwest valley deposit showing the location of the individual sections.

Plate 4.6 The distal part of the southwest-valley deposit, with the ice covered South Lake on the extreme right. Note the person on top of the deposit for scale.

Plate 4.7 Section 6 representing the most distal part of the southwest valley deposit, composed of ripple-laminated fine sands and silts.

bed accumulation.

A second erosional contact defines the upper part of the section, which consists of very fine sand in the form of type B, ripple-drift cross-lamination (subfacies 2, Plate 4.7). This stratification indicates that the currents were strong enough (40-45 cm/sec) to produce small ripples by bed traction, however, deposition from suspension was high enough to prevent erosion of the stoss sides. Occasionally, within this upper part of the section (subfacies 2), erosional surfaces are apparent, indicating that the stratification was transitional to type A, ripple-drift cross-lamination (subfacies 1). The fine sands, therefore, predominantly represent the complete profiles of ripples migrating with a large angle of climb.

Collectively, the Facies I sequence (section 6) is interpreted as pulsating underflows of relatively low velocity. The underflows fluctuated across the threshold velocity for bed traction of fine sands, and therefore changed the suspended to traction load ratio. This change was sufficient to produce three subfacies within the sequence, which represent relative changes from high to low to intermediate current velocities upward through the section. However, the presence of erosional contacts between the subfacies indicate that the velocities producing the subfacies did not grade from one into another.

4.2.2.2 Facies II - Planar-laminated sands and silts

Planar-laminated sands and silts comprise a much smaller part of the lacustrine deposit in southwest valley, usually occurring as 10 to 20 cm units interrupting a Facies I sequence. Two subfacies can be distinguished on the basis of grain size. Subfacies 1 is planar-laminated fine sand, which contains infrequent coarse sand/fine gravel beds. Subfacies 2 is planar-laminated silt, made up of 1 to 2 mm laminations.

Subfacies 1 comprises the top 8 cm of section 6 (Plate 4.7). The individual laminae are thin (2 to 10 mm) and composed of moderately well-sorted fine sand. They are interpreted as upper-stage plane beds, deposited from traction by turbidity currents too powerful for current ripple development (>70 cm/sec). Probably, these stronger currents

were related directly to higher discharges of the meltwater stream. The presence of fine gravels indicates that the currents were prone to occasional surges, which interjected this coarse sediment into the otherwise fine sand beds. Mass movements in more proximal (coarse-grained) parts of the deposit may have been responsible for the gravels.

An example of subfacies 2 occurs at section 4 (Plate 4.8), which is 80 cm high and represents the central part of the southwest valley deposit (Figure 4.3). The lower 60 cm of section 4 consists of a Facies I sequence composed of ripple cross-laminated, silty sands and organics. There is a sharp contact between the Facies I sediments and the overlying silts (Facies 2), suggesting that there was a sudden reduction in current velocity which brought about deposition entirely from suspension. It is concluded that the planar-laminated silts were deposited by slow-moving turbidity currents (<15 cm/sec), too weak to cause subsequent bed traction. The silts are very poorly sorted, suggesting that the finest component of the laminae (30% clay) settled out gradually in still-water conditions.

The southwest valley does not appear to have been an ice-contact environment during deposition of these sediments into South Lake. This part of the expanded lake was probably well beyond the ice margin, therefore the lacustrine deposits represent more distal, proglacial sedimentation. Hence, the southwest valley serves to contrast with the coarse, ice-contact delta on the east side of South Lake.

4.3 Plateau Lake

4.3.1 The Christie Section

The Christie section is the main lacustrine deposit in the study area, and it has been exposed by both the Divide and Plateau rivers (Figure 4.1). The deposit is 1.2 km long, covers 0.27 km^2 and its top surface is about 45 m above the confluence of the two rivers. A small western extension of the deposit, however, is about 13 m lower than the main surface.

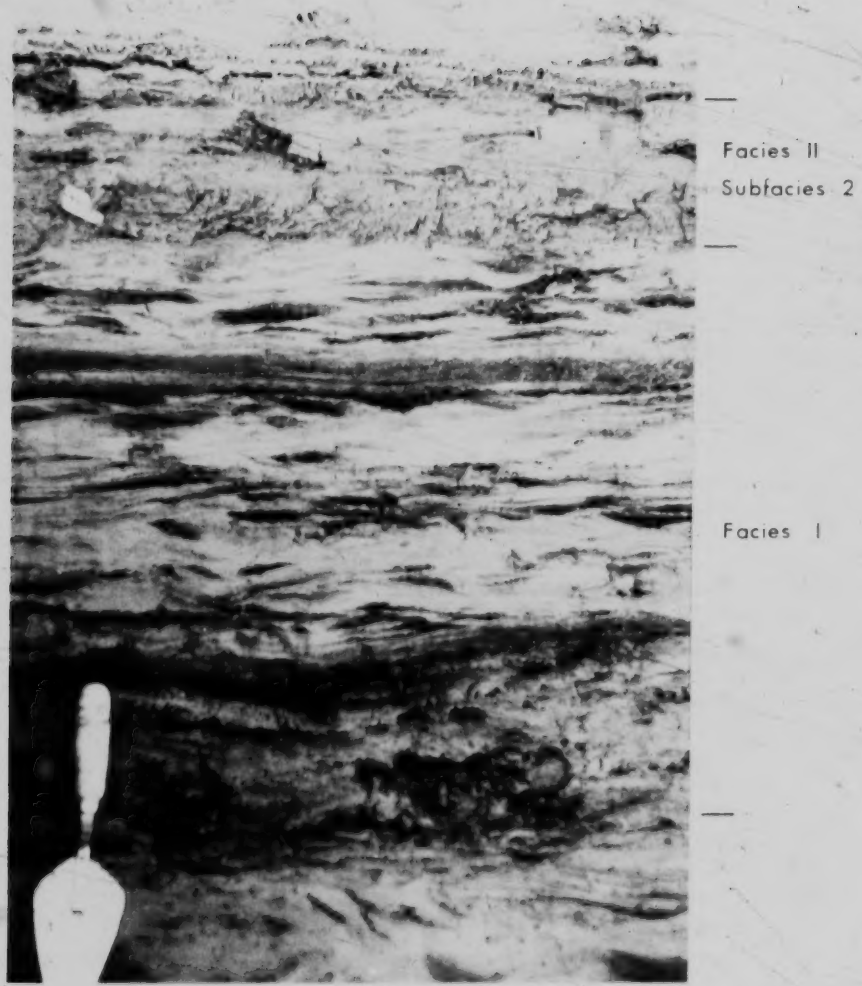

Plate 4.8 Section 4 from the central part of the southwest valley deposit, composed of Facies I sediments interrupted by planar-laminated silts marked by the key.

Vegetation partly conceals the sediments, and the degree of exposure varies with slope angle. For example, the Plateau river is not under-cutting the deposit, therefore slopes on the south side of the Christie section are stable at about 35°, and predominantly covered with vegetation (Plate 4.9). On the other hand, the more powerful Divide river has undercut the north side, resulting in slope angles as high as 75° at the main cliff (Plate 4.10). Mass movements on these steep slopes maintain clean exposures on the north side of the Christie section, however the resulting colluvium obscures the lower third of the section.

The Christie section shows a progressive fining of sediment in the distal direction (generally in the downstream direction of the Divide river). The distal half of the section is dominated by sediments smaller than medium sand. Organic matter is present throughout the deposit, and in places constitutes one-third of the volume (Christie, 1967). In proximal areas (toward the Divide glacier) coarse sands and gravels appear to be the dominant components, however, a greater amount of colluvium obscures the slopes in these areas. In addition, these slopes may be partially mantled by coarse sediments subsequently deposited by the Divide river. This makes it very difficult to identify *in situ* sediments, therefore only 9 of the 25 sections studied were located in the proximal half of the Christie section (Figure 4.4).

Four facies have been identified from the 25 lm sections. Two of the facies, ripple cross-laminated fine sands (Facies I) and planar-laminated sands and silts (Facies II), are equivalent to those reported from the southwest valley. One of the remaining facies represents higher energy conditions similar to the ice-contact delta (Facies III), whereas the final facies represents calmer conditions in a more distal environment (Facies IV). The facies of the Christie section will be presented as a progression from proximal (higher energy) areas to distal (lower energy) areas in the sequence III, II, I and IV. The main cliff is considered as the central part of the deposit (Plate 4.10, Figure 4.4) and individual sections are related to it.

4.3.1.1 Facies III - Cross-stratified sands and fine gravels

Facies III consists of coarse sediment deposited in a high energy environment. This facies occurs in two areas of the Christie section: 1) in proximal areas, east of the main

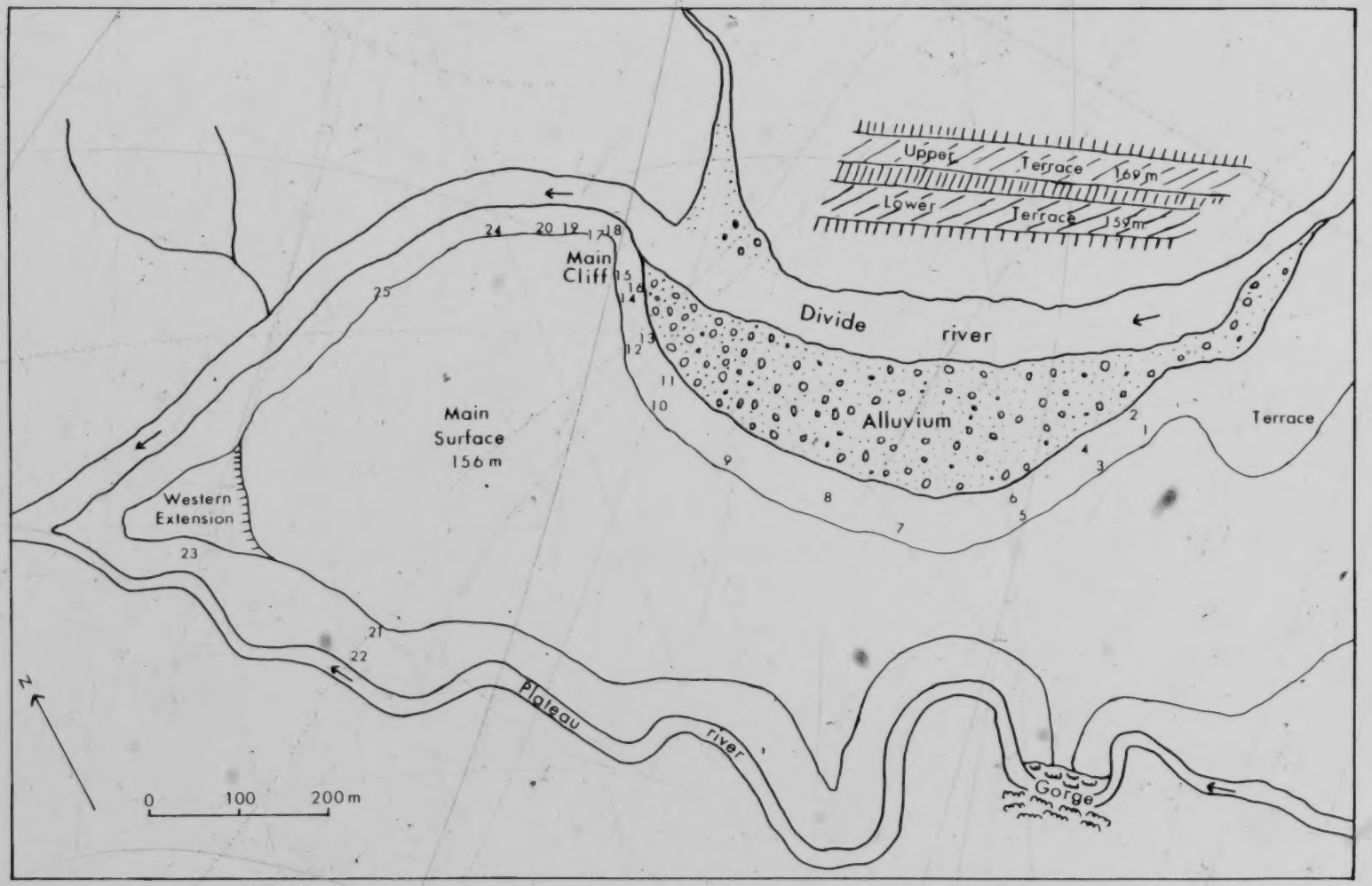

Figure 4.4 Field sketch of the Christie section, showing the location of the individual sections.

Plate 4.9 Looking northwest across the Christie section, the Plateau river and vegetated south side can be seen whereas the top of the main cliff is in shadow. Note the lower western extension of the section in left distance.

Plate 4.10 Looking west at the main cliff of the Christie section, the top being about 30 m above the modern alluvium from the Divide river in the foreground. From left to right represents proximal to distal.

cliff; and 2) in the top metre of the deposit.

Section 10 is 75 cm high and located 4 m below the surface, proximal to the main cliff (Figure 4.4). It is composed of several sets of cross-stratified sands with rare fine gravel lenses (Plate 4.11). The cross-stratified sets vary in mean size from fine to medium sand and are inclined at 5° to 26°. The absence of ripple-drift cross-lamination indicates that the currents were too powerful (>60 cm/sec) for small ripples to develop. Because the sands are close to the size threshold for dune development (0.2 to 0.3 mm), the steeply dipping (26°) sands could have been deposited by the migration of dunes to produce large-scale trough cross-stratification. Such cross-stratification represents the foreset laminae of dunes that were preserved following the migration of the next dune. The finer, gently dipping (5°) sands, to the right of the trowel (Plate 4.11), could be upper-stage plane beds which developed in place of the dunes due to the smaller grain size of this unit. The interchanging of bedforms is a possible explanation for the cross-cutting of the cross-stratified sets.

Section 17 is 65 cm high and represents the top 1 m of the deposit, located at the top of the main cliff (Figure 4.4). The section is composed of cross-stratified sands and fine gravels (Plate 4.12). The central 20 cm of the section is composed of medium sand in the form of subparallel sets of cross-laminae. These sets were cut by a channel, which was then filled with cross-stratified coarse sands, in turn covered by horizontal fine sands. The coarsest sediments are at the base of section 17, and consist of steeply dipping (25° to 30°) fine gravel foresets, which contain a large amount of organics. These organics (to the right of the trowel blade in Plate 4.12) were collected and ^{14}C dated at >33,000 BP (S-2633). These cross-stratified sediments were deposited by relatively high velocity currents, powerful enough to produce dunes. These dunes migrated at a low angle of climb, producing large-scale, trough cross-stratification, in which the cross-laminae of each set represent a small fraction of the dune foreset beds.

Plate 4.11 Section 10, composed of cross-stratified sands and fine gravels, located 4 m below the deposit surface and proximal to the main cliff.

Plate 4.12 Section 17, composed of cross-stratified sands and fine gravels, located at the top of the main cliff. Organics within fine gravels to the right of the trowel blade were ^{14}C dated at > 33,000 BP.

4.3.1.2 Facies II - Planar-laminated sands and silts

The plane beds of the Christie section can also be divided into two distinct subfacies, based on grain size. Planar-laminated fine/medium sand (subfacies 1) is common in areas proximal (east) to the main cliff (Figure 4.4). Section 12 is predominantly composed of planar-laminated, fine sands dipping at 18° (Plate 4.13). Individual beds vary in thickness from 0.3 to 2.5 cm; they are very poorly sorted and their mean grain size ranges from coarse silt to fine sand (Table 4.1). These characteristics suggest that the sands are upper-stage plane beds, deposited by turbidity currents too powerful for ripples to develop (>70 cm/sec, Table 4.2). The sediments are positively skewed, which indicates that the peak velocity, or main pulse of the currents, deposited the coarsest components. Finer sediments were deposited as the currents waned, therefore producing graded beds with a large grain-size distribution (poor sorting).

Planar-laminated silt (subfacies 2) occasionally interrupts Facies I (ripple cross-lamination), both within the main cliff, and in areas to the west (distal) of it. The change from Facies I to subfacies 2 (Facies II) can be abrupt, however, there is usually a gradation from sinusoidal ripple-lamination, to planar lamination and back to sinusoidal ripple-lamination. No erosional contacts were observed, therefore this sequence represents only a minor change in current conditions; but because the velocity temporarily fell below the threshold for bed traction of silt, there was a facies change from ripple lamination (Facies I) to planar lamination (Facies II). Again, the silts are poorly sorted and positively skewed, which suggests initial deposition from slow-moving turbidity currents (<15 cm/sec) followed by vertical settling of the finest sediment (clays) in the planar laminae.

4.3.1.3 Facies I - Ripple cross-laminated fine sands

Facies I is the most commonly occurring facies within the Christie section, and it dominates the central deposits of the main cliff (Figure 4.4). Section 16, a 90 cm sequence from near the base of the main cliff, shows two subfacies of Facies I. The sequence is

Plate 4.13 Section 12, composed of planar-laminated sands, located proximal to the main cliff. This is typical of Facies II.

composed of silty sands which alternate between type A and type B, ripple-drift cross-lamination (Plate 4.14). The two types of cross-lamination can be distinguished by their geometry and colour. Organics collected from 62 cm down section 16 (marked by the top of the trowel handle) were dated at 15,040 ± 625 BP (S-2632).

The type A subfacies, consists of small cosets of yellow, silty sand and organic detritus. The cosets are the partial ripple-foreset cross-laminae remaining after ripple migration. They vary in thickness, indicating that the rate of deposition from suspension varied during ripple migration. A 3 cm horizon of planar-laminated silty sand (Facies II), interrupts a type A subfacies, 54 cm down section 16. The cosets become less distinct on either side of the plane bed, indicating a transition zone from type A, ripple-drift cross-lamination to upper-stage plane beds. As the current velocity increased, the rate of deposition from suspension decreased, until all the silty sand was deposited from traction as plane beds. No intervening dunes developed because the sediment size remained finer than the size required for dunes (0.2 mm, Harms et al., 1982).

The type B subfacies consists of larger ripples, approximately 20 cm long and 6 cm high, which are composed of orange/brown, silty sand and organics. The majority of this subfacies appears to be transitional towards the type A subfacies. In fact, at the top of section 16 (Plate 4.14), complete ripple profiles are missing, indicating the transition to type A, ripple-drift cross-lamination in the orange/brown silty sands.

The two cross-laminated subfacies both have mean grain sizes of 5.3 phi, and are very poorly sorted. The whole sequence is considered to have been deposited by density underflows, with velocity variations only sufficient to change the type of ripple-drift cross-lamination. At one time the underflow apparently crossed a velocity threshold (ca. 80 cm/sec), and plane beds were deposited. This increase in velocity, however, was not sufficient to transport significantly coarser sediment. The paucity of breaks in section 16 suggests deposition by a more or less continuous, although slightly pulsating, density underflow, representing one event.

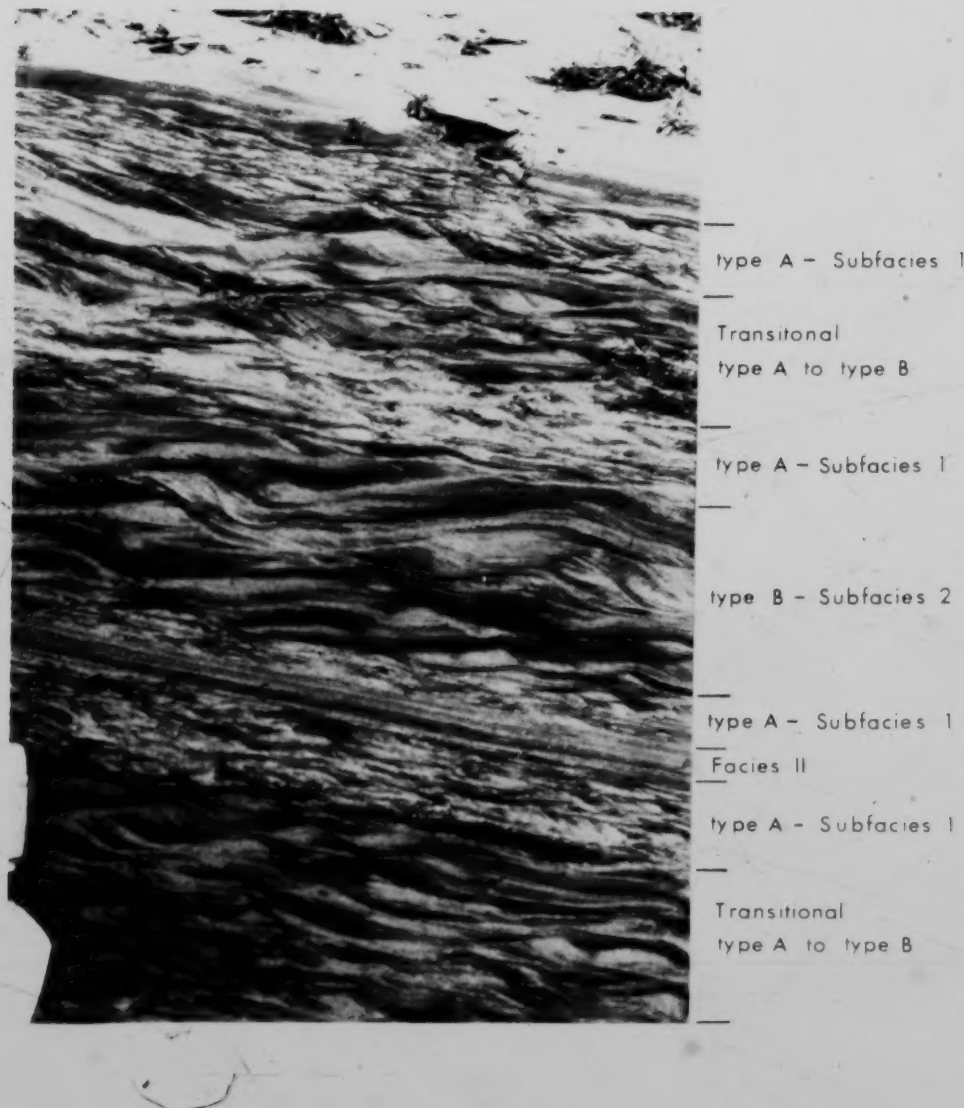

Plate 4.14 Section 16, composed of ripple-laminated silty sands, located near the base of the main cliff. Note, organic horizon at the top of the trowel handle which was ^{14}C dated at 15,040 ± 625 BP.

4.3.1.4 Facies IV - Rhythmically-bedded silts and clays

The planar-laminated silts and clays in section 22 (Plate 4.15) have been designated Facies IV, rather than a subfacies of Facies II, for two reasons. First, the silts and clays are rhythmically-bedded, alternating between brown silt laminae and blue/grey clay laminae. Second, the stratification is continuous throughout the section, with no apparent transition to sinusoidal ripple-lamination.

Section 22 is 75 cm high and located on the south side of the Christie section, 8 m above the Plateau river (Figure 4.4). It is only 200 m from the western extremity of the deposit, therefore it represents a distal location. The silt and clay laminae are frequently arranged in couplets; thicker, brown silt horizons (6-15 mm) and thinner, blue/grey clay horizons (2-5 mm), which could be equivalent to the Group 3 varves of Ashley (1975). Thin organic horizons (1 to 3 mm) sometimes occur within the silt laminae (Plate 4.16). The thickness of the laminae, as well as the presence of organics, indicates that the silts were deposited from slow-moving turbidity currents (<15 cm/sec). However, the finest sediments (clay) of each rhythmite probably settled vertically under tranquil conditions. The lack of micro-cross lamination indicates that the currents were not strong enough to transport the silt as bedload subsequent to deposition. Very thin (1 mm) yellow horizons (as seen near the top of Plate 4.16) show that occasional turbidity currents were strong enough to transport very fine sand in suspension. The clay laminae probably accumulated by vertical settling, when turbidity currents were not prevalent. Overall these sediments represent distal, lake floor sedimentation, and the absence of ice-rafted clasts suggests permanent lake ice.

The Christie section contains the greatest range of lacustrine sediments within the study area. It provides a good example of a sedimentary progression from the margin, to the floor of a lake, that was close to an ice source. Initially, coarse sediments (gravels) were deposited rapidly at the point of meltwater entry into the former lake. With increasing distance from this point, progressively finer sediments were deposited, until

Plate 4.15 Section 22, located near the base on the south side of the Christie section. The key marks thin organic and yellow fine sand horizons which occasionally occur within the silt-clay rhythmites.

Plate 4.16 A close-up of part of section 22. Note the thin organic horizons above the coin and the yellow fine sand horizon near the top.

clays settled-out on the lake floor. This change in sediment size reflects the waning power, as well as the increasing interaction, of the meltwater currents in the ambient water.

4.3.2 The Central Valley Section

A smaller lacustrine deposit has been exposed by the Central valley stream, 1.1 km south of the Divide valley (Figure 4.1). The section is about 180 m wide and 21 m high, however, the sediments are partly concealed by vegetation and colluvium at the base (Plate 4.17). The sedimentology of the deposit was studied from eleven, 1 m sections cut into the exposed sediments. Generally, the deposit is finer grained than the Christie section, however, there is no apparent fining trend in the distal direction (northward towards Piper Pass). Only two facies were recognized: ripple cross-laminated fine sands (Facies I), and planar-laminated sands and silts (Facies II). Facies II dominates the stratification, with Facies I appearing only in the three top sections (Figure 4.5).

4.3.2.1 Facies I - Ripple cross-laminated fine sands

Facies I is poorly represented compared to its dominance within the southwest valley and Christie sections. This is shown by the absence of both type A and type B, ripple-drift cross-lamination (subfacies 1 and 2). Sinusoidal ripple-lamination is the only subfacies of Facies I that occurs at the Central valley section, although in places, it becomes transitional to type B, ripple-drift cross-lamination.

Section 6 is 95 cm high and located at the top of the Central valley section (Figure 4.5). It is composed of fine sands and silts in the form of planar lamination and sinusoidal ripple-lamination (subfacies 3) (Plate 4.18). Organics collected from 78 cm down section 6 were dated at 19,055 ± 2930 BP (S-2635). The size and stratification suggests that the silty sands were deposited by relatively low velocity underflows, which fluctuated across the threshold for bed traction (ca. 20 cm/sec). Consequently the planar laminae were deposited from suspension by low velocity underflows, and the sinusoidal ripple-lamination was deposited from a combination of traction and suspension at higher

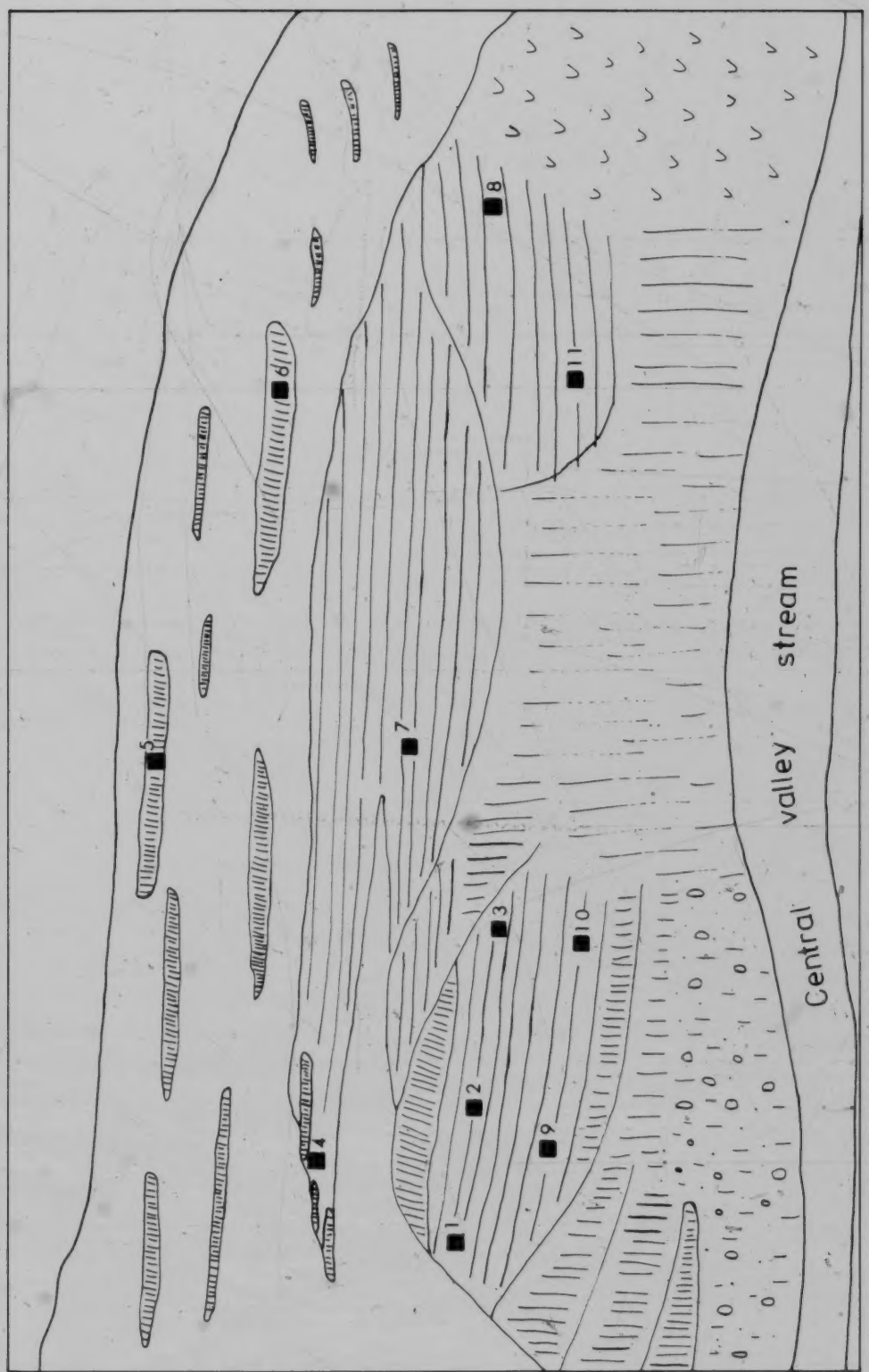

Figure 4.5 Field sketch of the Central valley section, showing the location of the individual sections.

Plate 4.17 Looking east at the Central valley section which reaches a height of about 20 m above the stream. Piper Pass is to the right.

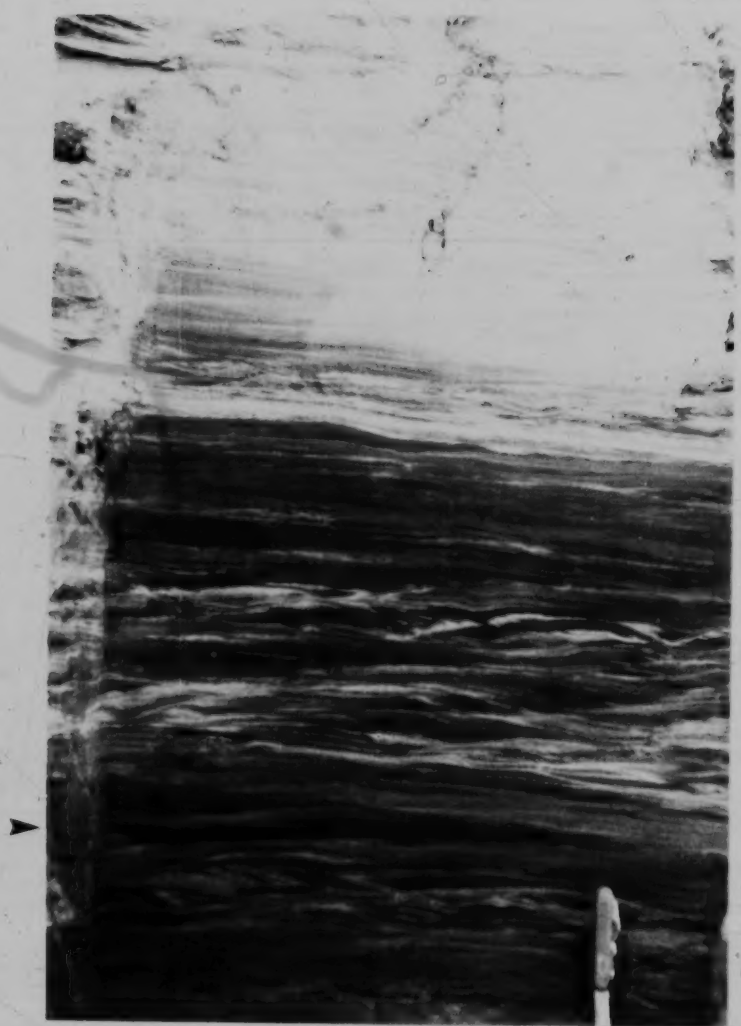

Plate 4.18 Section 6, composed of silts and fine sands, located at the top of the Central valley section. Arrow indicates horizon from which organics were ^{14}C dated at 19,055 ± 2930 BP.

velocities. The absence of type A and type B, ripple-drift cross-lamination (subfacies 1 and 2), indicates that deposition from suspension remained dominant throughout, and the currents were not able to produce complete ripples. The ripple cross-laminated, silty sands are very poorly sorted and postively skewed due to a 20% clay component. This suggests that finer sediments settled-out during the waning of each underflow, however no clay drapes are visible. Therefore it is likely that the fines were then incorporated into ripple cross-lamination produced by the next underflow.

4.3.2.2 Facies II - Planar-laminated sands and silts

Facies II dominates the sedimentology and shows the same characteristics throughout the deposit. Section 3 is 90 cm high and located in the middle of the overall section (Figure 4.5). It is composed of silt and clay (subfacies 1), however, light coloured sand horizons interupt the sequence (Plate 4.19). The planar-laminated silts and clays were depostied entirely from suspension by slow-moving turbidity currents (<15 cm/sec). Fine sand horizons, for example near the bottom of Plate 4.19, can range from a few millimeters to 8 cm. These sands are fine enough to have also been deposited from suspension by slow-moving currents. An alternative interpretation is that the fine sands were deposited by short-lived, powerful turbidity currents (>70 cm/sec, Table 4.2). These currents must have maintained their velocity throughout deposition because ripple cross-lamination, indicative of waning power and deposition from suspension, is not present within the sands. The moderate sorting of the thicker sand horizons suggests that they are upper-stage, plane beds deposited from traction, although there is no evidence of scouring or rip-up of the underlying beds. The thin (0.5 cm) sand horizons were probably deposited entirely from suspension.

The Central valley section is composed of sediments finer than medium sand, therefore it represents either sedimentation distal to the meltwater input (i.e. equivalent to the western end of the Christie section), or sedimentation by low energy streams. Deposition from suspension dominated at the site, however, powerful turbidity currents

Plate 4.19 Section 3, composed of planar-laminated silts and sands, located in the middle of the Central valley section.

may have occasionally interrupted the calm conditions to deposit thick beds of fine sand. Isolated gravel clasts sometimes occur within the plane beds, representing ice-rafted dropstones probably deposited from lake ice. A second organic sample, collected from section 11 near the base (Figure 4.5), was dated at 20,950 ± 1200 BP (S-2634).

4.4 Summary

Examining the stratigraphy and sedimentology of many individual sections throughout the field area indicates that the four sites represent environments at different distances from former ice margins. The ice-contact delta is a recent example of lacustrine sedimentation in contact with a glacier. On the other hand, sediments deposited in the southwest valley provide a good example of non ice-contact proglacial sedimentation. Therefore, the lowering of South Lake allows the detailed examination of a wide range of glaciolacustrine sediments deposited into a known environment (Figure 4.1).

The glaciolacustrine deposits south of Piper Pass reflect sedimentation in different areas of a former lake. The Christie section, with its complete range of sedimentology, represents a progression from proximal to distal sediments; whereas the generally fine-grained Central valley section represents only distal sedimentation (Figure 4.1). The sedimentology between the four sites varies in grain size and stratification, however the mode of deposition, by turbidity currents of varying strengths, appears to be similar from one site to the next.

The deposits within the study area reflect the likely proximity to ice. The study area is located midway between two previously studied areas, Clements Markham Inlet (Bednarski, 1984, in press; England and Bednarski, in press) and Archer Fiord/Lady Franklin Bay (England, 1983; Figure 1.1), where limited ice during the last glaciation has been documented. Interpretation of the stratigraphy and sedimentology combined with the ^{14}C dates, will elucidate the paleogeography presented in the next chapter.

5. PALEOGEOGRAPHY AND CHRONOLOGY

Examination of the glaciolacustrine sediments within the Piper Pass area indicate the style of former meltwater deposition. These deposits, in turn, indicate the proximity of local ice margins to the lake environments. Therefore, the paleogeography of South and Plateau lakes is reconstructed on the basis of the stratigraphy and sedimentology of their glaciolacustrine sediments, and the relationships of these sediments to additional lake and glacial deposits. The ^{14}C dates provide a chronological framework for South and Plateau lakes, as well as for past glacial activity in the study area.

5.1 South Lake

5.1.1 Ice-contact delta

The prominent delta, recently constructed during retreat of the valley glacier, displays an internal composition reflecting local changes in the bedrock. The glacier, aligned along the northern extension of the Hazen Thrust Fault, is actively eroding different bedrock types along its north and south flanks (Figure 1.3). North of the glacier, blue-grey shales of the Grantland Formation are being eroded, in contrast to yellow Permian sandstones south of the glacier. This difference in erosion is shown at the confluence of the streams exiting either side of the glacier (Plate 4.3). For example, the south stream transports yellow-brown sandstones which overwhelm the blue-grey coloured, north stream below their confluence. During previous higher lake levels these two streams entered South Lake independently. Consequently, the south half of the ice-contact delta is composed mainly of yellow sandstone clasts, whereas the north half is mainly composed of blue-grey shale clasts.

This contrast in delta composition indicates that construction of the ice-contact delta was controlled by both meltwater streams. They were powerful enough to transport clasts up to boulder-size and deposit these coarse sediments onto stagnant ice as the glacier retreated from South Lake. The delta prograded into the lake by underflows avalanching across this remnant

ice, depositing steeply dipping foreset beds. The varying composition of individual foresets reflects discharges that likely changed over intervals of a few hours to a few days. During small discharges, or in areas marginal to the meltwater streams, sands and silts were deposited by less powerful underflows.

The highest terraces (T4) of the ice-contact delta are accordant with the uppermost shoreline of South Lake. This indicates that the glacier had receded as far as the T4 terraces before the lake began to lower. The ^{14}C date of 1585 ± 190 BP (S-2637) from section 20 (Plate 4.5) within T4S (Figure 5.1), indicates that the lake level has fallen 29 m over the past 1600 years. The development of a delta staircase consisting of four terrace levels indicates that South Lake descended in four stages. However, a possible one metre subhorizontal bed of topsets interrupting the foreset beds of T2N (Plate 4.3) suggest that the fall in lake level was more complicated. If the darker coloured horizon within T2N is a topset bed, rather than sediment containing a difference in ice content, then this implies a 6 m transgression occurred within the overall lowering of South Lake and a second couplet of foresets and topsets were deposited. The glacier continued to recede as South Lake lowered, losing contact with the lake following the formation of the T2 terrace (16m above lake level). This is indicated by the formation of T2N(M), which is the most proximal terrace surface and an extension of a moraine ridge. This ridge must have been part of the glacier margin and therefore protected from the planing effects of the lake (Figure 5.1).

The different rates at which South Lake descended from the uppermost shoreline are unknown. However, some recent changes can be determined by reference to aerial photographs of the study area taken in 1958. At that time, the two meltwater streams were not actively tunnelling beneath the delta terraces, and they occupied channels that have since been abandoned. In addition, the T1 terraces were not exposed, and the lake level was 12 to 13 m higher than present (Plate 5.1). Therefore, South Lake has lowered approximately 13 m during the last 26 years or less.

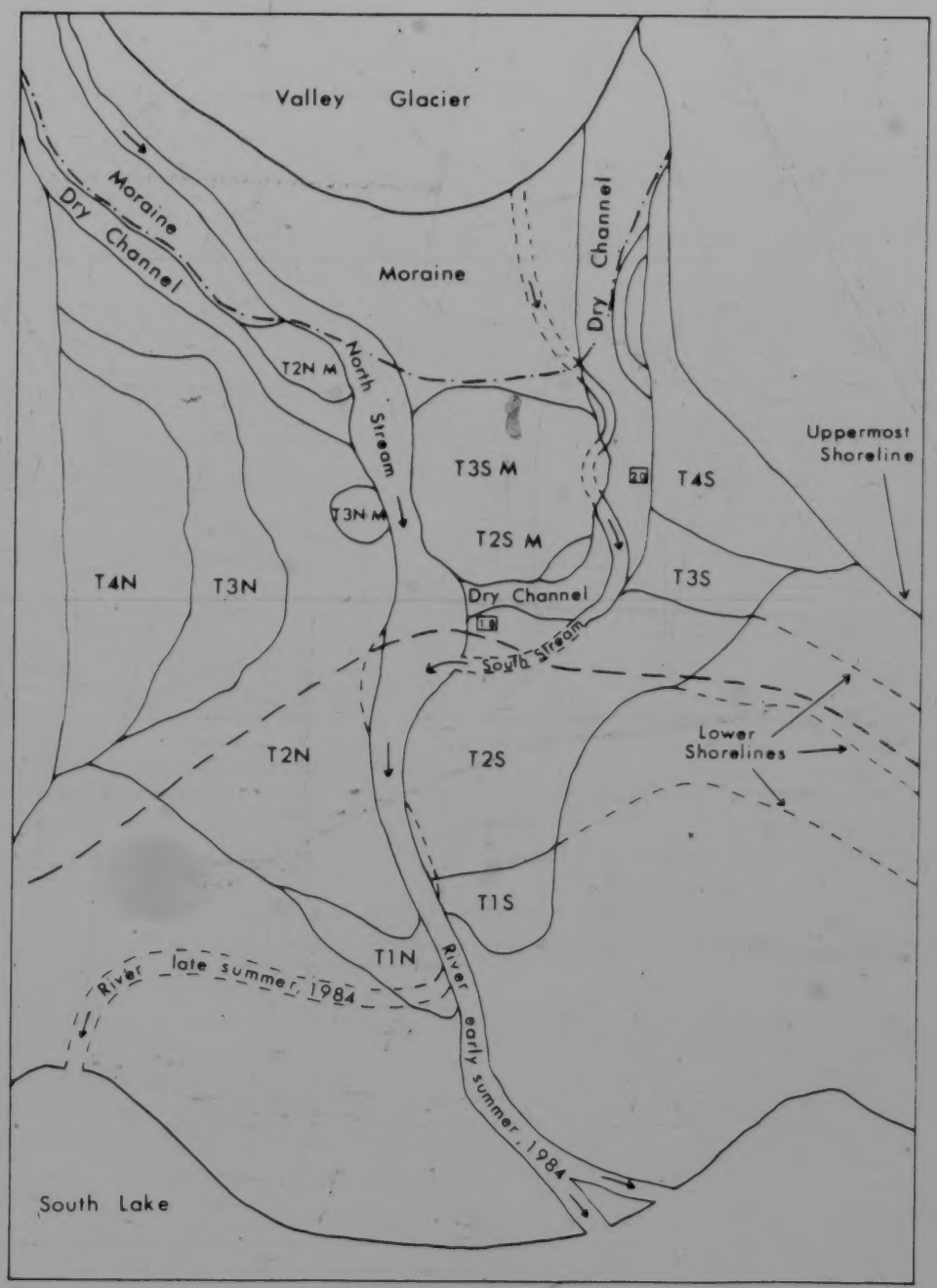

Figure 5.1 Field sketch of the ice-contact delta showing the relative positions of the terraces and meltwater channels. Also shown are previous positions of South Lake and the glacier.

Plate 5.1 The ice-contact delta from the air, enlarged from a 1958 aerial photograph. Note the prominent T4N and T3N terraces, however T2 is just forming and T1 is not exposed. Compare with Plate 4.1 and Figure 5.1.

At present, the south meltwater stream flows beneath the glacier snout and exits from T3S(M). This is important for the continued redistribution of sediment by thermal erosion and undercutting of the south side terraces. The combined effect of the meltwater streams in modifying the delta can be catastrophic. For example, during the fieldwork, the river broke through T1N and completely changed its course, entering the lake at the north side of the delta (Plate 5.2). Further modification of the delta's structure would be enhanced by continued lowering of the lake.

A similar study on glaciomarine deltas has been conducted at the head of Clements Markham Inlet (Figure 1.1), and these deposits contrast with the ice-contact delta in this study (Stewart, in preparation). Initially, an ice-contact marine delta was constructed at 103 m asl, approximately 7,800 BP (Bednarski, 1984). Subsequent sea level regression caused by postglacial emergence led to the formation of a delta staircase, with additional terraces at 70, 50 and 33 m asl. The stratigraphy of these deltas is poor, with no evidence of foreset beds. However, this is not attributed to the inclusion and subsequent thermal erosion of remnant ice within the delta complex. Throughout deposition, many mass movements at the delta front are considered to have redeposited coarse sediments in the form of deep water fans. Therefore, the only recognizable deltaic stratigraphy in the complex are silt and clay bottomset beds (Stewart, in preparation).

The raised marine deltas of Clements Markham Inlet provide a marked contrast to the entirely ice-cored, Gilbert-type delta examined on the east side of South Lake, which contains pronounced foreset and topset beds. The youthfulness of this ice-cored delta explains its well-preserved stratigraphy. However, future erosion of the ice core will continually reduce the size of the delta, and given the changes observed during the 1984 field season it could be entirely removed in the next 20-30 years.

Plate 5.2 Looking east at the distal end of T1N, which has been cut through and partially collapsed by the river.

5.1.2 The Southwest Valley

The confined southeast and southwest valleys, occupied by the expanded South Lake, were also sites of sedimentation during construction of the ice-contact delta. At its highest stage, South Lake extended 2.5 km up the southwest valley, and sediments accumulated up to at least 16 m near the valley mouth. Ripple cross-laminated fine sands (Facies I) dominate the sediments, and indicate that sedimentation was controlled by underflows (ca. 20-60 cm/sec) which fluctuated both temporally and spatially.

The overall sedimentology of the southwest valley deposit illustrates temporal variations more clearly than spatial variations in underflow velocities. At any one section, different subfacies of ripple cross-laminated sands can be identified, which reflect changes in current velocity over time. The change from sinusoidal ripple-lamination (subfacies 3) through to type A, ripple-drift cross-lamination (subfacies 1) represents faster currents capable of not only transporting coarser sediments, but of depositing a greater proportion from traction instead of suspension (Table 5.1). The poor sorting and positive skewness of some individual beds suggest that single currents also fluctuated over time, most likely waning and depositing progressively finer sediment. More significant velocity changes are indicated by the interruption of a Facies II unit within a Facies I sequence. The very-poorly sorted, planar-laminated silts (Facies II_2) were probably deposited by a combination of slow-moving turbidity currents and vertical settling (27% clay component, Table 5.1).

A Facies I sequence consisting of sinusoidal ripple lamination and ripple-drift cross-lamination, grading upward into planar lamination, may be interpreted as the C and D divisions of the Bouma sequence (Bouma, 1962). This interpretation implies that the sequence was deposited from one event. A high density turbidity current responsible for depositing a large volume of sediment can be generated by either a high velocity flow capable of entraining a large amount of sediment, or by a large slump.

In the arid environment of the High Arctic, high discharge/velocity events are predominantly generated by glacial meltwater. At the southwest valley and Central valley sites,

TABLE 5.1

Southwest Valley - 6 Sections, 12 Samples

Facies	Subfacies	Graphic Mean ø	Inclusive Graphic Standard Deviation ø (Sorting)	Percentages Sand	Silt	Clay	Section #
I	1	3.15	0.58	92	8		3
Ripple cross-laminated silty sand	1	3.18	0.68	90	10		6
	2 transitional to 1	2.95	0.51	94	6		5
	2(?)	3.70	1.03	78	22		1
	2(?)	3.55	1.19	77	23		2
	2	3.97	1.25	71	29		6
	3 transitional to 2	4.28	1.75	60	29.5	10.5	6
	3	4.47	1.84	57	31	12	3
	3	4.43	1.79	52	35.5	12.5	4
	3	4.15	1.62	63	26	11	5
II							
Planar laminated sands and silts	1	2.58	0.54	96	4		3
	2	7.27	2.40	0.1	73	26.9	4

TABLE 5.1(a)

Verbal classification scale for sorting (Folk, 1974 p. 46)

under .35ø, very well sorted	1.0-2.0ø, poorly sorted
.35-.50ø, well sorted	2.0-4.0ø, very poorly sorted
.50-.71ø, moderately well sorted	over 4.0ø, extremely poorly sorted
.71-1.0ø, moderately sorted	

there is no evidence of a former proximal ice margin that could generate a large amount of meltwater capable of creating a high density turbidity current. Neither are there extensive sands and silts upstream of these sites that could be susceptable to large scale slumping. A final point is that in many of the observed Facies I sequences, erosional contacts separate the different subfacies, indicating that they were not deposited by one turbidity current. The above points argue against a Bouma sequence origin for the Facies I and II sequences, and support the deposition of individual beds from discrete, low density turbidity currents.

The southwest valley deposits are sufficiently extensive (2.5 km) that a noticeable fining trend in the distal direction should be expected. In proximal areas the sediments are generally coarser, however there is no indication that an ice-contact environment ever existed during sedimentation. Present glaciers occur about 11 km upvalley, however there is no evidence of their last ice limit in the valley. A significant quantity of sands and fine gravels was derived from a tributary stream on the west side of the valley, 1.6 km from the present shore of South Lake. This steep gully constructed a small delta 26 m above South Lake and up to 12 m above the valley floor (Figure 4.3). Subsequent lowering of the lake, however, led to considerable modification of this delta by the main river. Additional glaciofluvial outwash contributes to the coarse appearance of the sediments in more proximal areas.

Sediments near the valley mouth do not show the expected characteristics of distal sedimentation. At distances greater than about 1.5 km from the meltwater input, silt and clay laminae, possibly interrupted by occasional sand horizons, normally dominate a glaciolacustrine succession (Ashley, 1975; Shaw, 1977; Shaw and Archer, 1978). However, in this deposit, the central and distal parts are very similar; laminae are absent and the succession is dominated by ripple cross-laminated silty sands (e.g. Section 6, Plate 4.7). Factors controlling the depositional environment within southwest valley include changes in both the local ice margins and lake depth. These two variables can change the energy of the environment, and hence alter the mode of sedimentation, particularly under the influence of a proximal ice margin. A retreating ice margin generates large quantities of meltwater which, coupled with a decrease in

water depth (due to lake lowering and/or sediment accumulation), produces a higher energy environment. On the other hand, an ice advance generates less meltwater which, combined with a rise in lake level, produces a lower energy ('distal-type') environment.

Organic material collected from section 5, located about 6 m below the deposit surface (Figure 4.3), has been dated at 495 ± 170 BP (S-2636). The southwest valley sediments rise only to 16 m above South Lake, therefore this organic sample relates to a former lake level ≥ 16 m. Consequently, the maximum date for this shoreline, both stratigraphically and radiometrically, is 495 BP (because of redeposition of possibly older material). At present, the ice margin at the head of southwest valley is 8.5 km from the closest lacustrine sediments and 11 km from South Lake (Figure 1.2). Considering the recency of the deposits, it is unlikely that changes of this ice margin significantly affected the depositional environment near the valley mouth. Therefore, changes in lake elevation are considered the primary controls during this sedimentation.

Formation of the four terrace levels at the ice-contact delta indicates that there were at least four periods of lake level stability. Furthermore, numerous faint strandlines on the lake's southeast shore, together with the evidence from the aerial photographs, suggest that periods of lake lowering were not rapid events. The apparent 13 m fall in lake level at the ice-contact delta since 1958, is supported by South Lake having receded about 1 km from the southeast valley, and 600 m from southwest valley during the last 26 years. Based on this, it is conceivable that South Lake rose to the uppermost shoreline at a rate equal to or lower than the recent 50 cm/year change. Therefore, sediment accumulation within the confined southwest valley may have been rapid enough to prevent a deep-water, 'distal-type' environment from developing. The persistence of underflows in a relatively shallow environment would explain the dominance of ripple cross-laminated sediments and the similarity of the southwest valley deposit. Sediment accumulation would have continued until South Lake declined to the height of the deposit. Subsequent meltwater, however, was not powerful enough to cap these sediments with extensive outwash gravels.

5.1.3 The Uppermost Shoreline

5.1.3.1 Differential Tilting

Previous studies in Clements Markham Inlet (Stewart and England, 1983; Bednarski, 1984, in press) and Archer Fiord/Lady Franklin Bay (England, 1983) presented relative sea-level curves which indicate the amount of crustal emergence since the onset of glacial unloading (ca. 8,000 - 10,000 BP). These data indicate that tilting caused by differential emergence is measureable over a short distance (<10 km) provided that the shoreline is several thousand years old, because it is during that earlier interval that emergence is most rapid. In Piper Pass the uppermost shoreline is the oldest and best preserved shoreline above South Lake. Prior to, and since its formation, the study area has experienced continuous postglacial emergence and may have resulted in tilting of this shoreline. The detailed surveying, however, did not reveal any observable tilt. The shoreline's elevation above South Lake remained between 27 and 28 m along most of the 6 km distance (Figure 5.2). The additional altimeter results along both sides of South Lake, north to the damming glacier, also did not reflect tilting. Furthermore, many of the altimeter readings were outside the error range of ± 2m at this height (28 m) compared to a Quickset level. This was also compounded by the difficulties involved in pin-pointing the uppermost shoreline on the steep, scree covered slopes in this northern part of South Lake.

The ^{14}C dates demonstrate that the uppermost shoreline of South Lake is no more than 1600 years old. The lack of tilt is understandable because during the past 1600 years the rate of emergence has slowed down such that differential tilting along a 6 km shoreline would be negligible. Finally, the surveying results confirm that the expanded South Lake (+28 m), extending towards the Hazen Plateau, was not responsible for the higher lacustrine deposits within the Divide and Central valleys. The southeast valley must have been blocked at a previous time, and this led to the formation of a separate proglacial lake (Plateau Lake) south of Piper Pass.

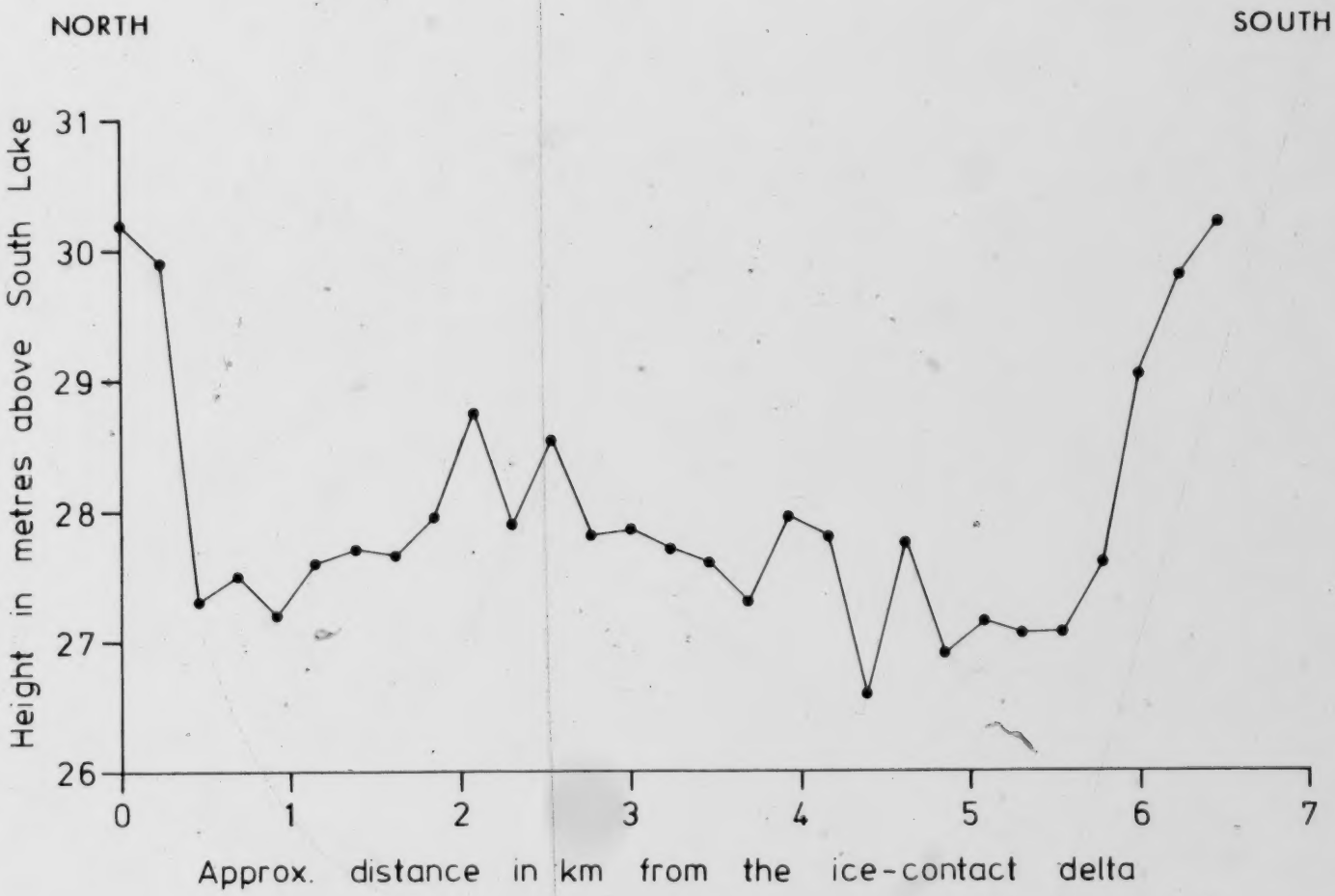

Figure 5.2 The elevation changes of the uppermost shoreline above South Lake, surveyed in detail from the ice-contact delta into the southeast valley

5.1.3.2 History of Shoreline Regression

The ^{14}C date from T4S of the ice-contact delta, dates the uppermost shoreline at $\leq$ 1600 BP. The sediments related to a shoreline of $\geq$ 16 m within the southwest valley date about 500 BP. However, this 500 BP date does not necessarily relate to the uppermost shoreline, therefore initial drop from 28 m could still be 1600 BP. On the other hand, it must be stressed that these dates provide only maximum ages for the shorelines, and that the aerial photographs suggest that the lake fell 13 m since 1958. Therefore, the following discussion concerning control of these changing lake levels should include any recent events that may be of significance. However, the considerable time lag between climatic change and the response of high latitude glaciers, necessitates the consideration of events prior to 1600 BP that may have influenced the lake's history.

5.1.4 Discussion

The most important control over the elevation of South Lake is the glacier damming its north end. At present, a 20-m trough between this glacier and the valley side allows drainage from South Lake into the central lake of Piper Pass (Figure 1.2). Minor fluctuations of this glacier margin would be sufficient either to restrict or increase the drainage from South Lake, thereby affecting its elevation. Glacier fluctuations of tens of metres could be caused by a variety of conditions ranging from a local response to the glacier profile or to regional climatic changes. The glacier damming South Lake descends the steep, west side of Piper Pass from the United States Range and terminates in an aqueous environment. These two factors facilitate small changes in the ice front in the absence of climatic change. In addition, fluctuations of the two other Piper Pass glaciers, descending from the Grant Ice Cap to the east, also have changed the northward drainage to Clements Markham Inlet (Figure 1.2). The restriction of this northward drainage, leading to a rise of lake elevation, is demonstrated by similar higher shorelines above the central and north lakes within the Pass (Figure 1.2). The three glaciers of Piper Pass may have responded independently, therefore it is not known whether the various

changes in lake elevations occurred synchronously or at different times.

The most significant climatic change in the High Arctic during the last few hundred years was the Little Ice Age. Many studies of arctic ice cores (Dansgaard et al., 1971; Koerner, 1977; Paterson et al., 1977; Fisher and Koerner, 1980) indicate that the substantial cooling of the Little Ice Age occurred from about 100 to 400 BP and that this was followed by a prominent amelioration most noteably during the first half of this century. It is conceivable that the glacier damming South Lake advanced in response to the Little Ice Age and closed the present trough allowing its drainage. An additional response to the cooler temperature would have been a smaller volume of meltwater entering South Lake, causing a slow rise to the uppermost shoreline.

These climatic fluctuations of the past 400 years are not the only possible control for the changing lake levels. The formation or expansion of the existing ice shelves along the north coast of Ellesmere Island is the most direct evidence for a climatic deterioration since ca. 4500 BP. Another indication of a late Holocene climatic deterioration in the High Arctic is that many outlet glaciers are at, or are retreating from, their maximum postglacial positions (Stewart and England, 1983). Therefore the expansion of the glacier that dammed South Lake to +28 m may have resulted from a climatic deterioration older than the Little Ice Age.

The increased temperatures at the beginning of this century possibly caused simultaneous retreat of both the damming glacier, and the valley glacier on the east side of South Lake. This would have lowered the lake elevation, as well as led to construction and subsequent exposure of the ice-contact delta. The continuation of lake level fluctuations is shown by its drop of 13 m during the past 26 years. Additional support for recent changes in the lake is provided by the organic sample from southwest valley dated at 495 $\pm$ 170 BP (S-2636) which relates to a shoreline $\geq$ 16 m. This sample contained mosses that were still green, indicative of a very young age (Lafarge-England, personal communication, 1985). This increases the possibility that the dated samples from South Lake provide maximum ages because they could well be based on a blend of organic material, containing fragments from older peat

deposits known to occur within the drainage basin (Lafarge-England, in preparation).

5.2 Plateau Lake

5.2.1 The Christie Section

The Christie section is the largest of the two glaciolacustrine deposits marking the former existence of Plateau Lake. Its location, at the confluence of the Divide and Plateau rivers, implies that the main sediment source for this part of Plateau Lake was the Divide Glacier. The Christie section contains the greatest sediment range of the four principal sites examined, and this range reflects the progression of sediment-laden meltwater currents through the lake. The stratigraphy and sedimentology indicate that the former meltwater streams entered Plateau Lake predominantly as underflows. These underflows varied temporally, and unlike the southwest valley sediments, a pronounced waning of current is also indicated in the distal direction.

The Christie section is considered to be a prograding delta. Coarse sands and gravels were deposited in proximal parts, and gently dipping (about 10°) sandy foreset beds prograded across silt and clay bottomset beds in more distal parts (Clemmensen and Houmark-Nielsen, 1981). The present slope of Divide valley is about 0.02 (1 m in 50 m) to the west, therefore the gently dipping sediments are partly a reflection of the corresponding former lake floor. It is not known whether an ice margin was ever in contact with Plateau Lake, but ice-marginal deposits east of the section do suggest that, at one time, Divide Glacier terminated at least 0.6 km from the lake basin (Plate 5.3).

The east end of the Christie section is mainly concealed by colluvium, therefore a detailed comparison was not possible between these most proximal sediments and the ice-contact delta built into South Lake. However, from the limited exposures, the sediments appear to consist of a coarse sand/gravel matrix, containing occasional cobble- and boulder-sized clasts. Some of the observed clasts may have been deposited as glacial outwash by the

Plate 5.3 Looking east from the top of the Christie section at the Divide river. Note, two terraces which mark former levels of Plateau Lake, and just beyond, irregular mounds which mark a former ice margin of Divide Glacier (black arrows).

Divide river. There are no steeply dipping beds, and the only recognized sorting within the matrix was crude bands of generally finer, and generally coarser gravels. The lack of steeply dipping beds, together with the relatively few large clasts, argues against ice-contact deposition and conforms to rapid deposition from powerful meltwater streams entering a relatively shallow lake environment.

Initially, the currents that flowed through Plateau Lake carried a wide range of sediments up to fine gravel size. Underflow velocities were sufficient to create dunes (>40-50 cm/sec), which migrated down the delta front and produced large-scale cross-stratification (Facies III). At times, the underflows were powerful enough to deposit all the sediments from traction (> 120 cm/sec), which prevented any ripple development. As a result, upper-stage plane-beds (Facies II), also frequent the proximal half of the Christie section. In addition to Facies II_1 and III (section 10, Plate 4.11), the occurrence of ripple cross-lamination (Facies I, e.g. section 11) within the same vertical sequence (Figure 4.4) demonstrates the significant temporal variations in underflow velocities. Furthermore, the poor sorting of these coarser facies indicates changing velocities over short time scales, probably due to waning flows and the incorporation of fine sediment into the beds (Table 5.2). The clay component increases the sorting of these beds, however, the calculated percentage of clay may be artificially high due to the inadequacies of the hydrometer analysis (Chapter 3).

With increasing distance from the point of meltwater entry, progressively finer sediment was carried by the waning underflows. The combination of sediment size (<0.25 mm) and current velocities (<60 cm/sec) decreased below the threshold for dune development, therefore small ripples migrated across the gently dipping delta front. The dominance of ripple cross-lamination (Facies I) in the vicinity of the main cliff, indicates that the underflows decreased in competence below this threshold after travelling a minimum of 650 m through the lake. The vertical persistence of the cross-laminated silty sands signifies the continuance of these underflows. However, the interbedding of the three subfacies (Facies I), as well as the occasional interruption of upper-stage plane beds, reflects their pulsating nature. On even

TABLE 5.2

Christie Section - 15 Sections, 32 Samples

Facies	Subfacies	Graphic Mean ø	Inclusive Graphic Standard Deviation ø (Sorting)	Percentages Sand	Silt	Clay	Section #
III		2.33	1.66	84	16		10
Cross		4.6	2.42	75	10	15	10
stratified		5.22	2.92	65	15	20	10
sands		0.53	1.54	94	6		17
		1.97	0.69	95.6	4.4		17
		1.95	0.98	93	7		25
II	1	4.8	3.47	65	15	20	12
Planar	1(?)	6.18	3.27	33	37	30	12
laminated	1	1.52	2.08	87	13		13
sands and	1	5.25	2.96	64	16	20	16
silts							
	2	6.55	2.80	19	53.5	27.5	14
	2	6.87	2.91	15.5	54	30.5	21
I	1	1.9	1.09	92	8		11
Ripple	1	2.67	1.43	83	17		11
cross-laminated	1	3.4	1.08	81	19		11
silty sand	1	5.35	2.81	64	17	19	16
	1	3.13	0.9	85	15		25
	2	4.37	2.36	73	12.5	14.5	14
	2	5.32	2.60	60	21.5	18.5	16
	2	4.75	2.55	68	16	16	20
	2	5.33	2.86	60	20	20	21
	2 transitional to 3	5.15	2.83	65	17	18	15
	2 transitional to 3	4.78	2.55	70	14	16	18
	3	5.93	2.91	45	27.5	27.5	15
	3	7.15	2.69	8	59	33	18
	3	5.32	2.69	56	25	19	19
	3	6.32	3.03	30	40	30	19
	3	5.63	2.70	47	33	20	20
	3	6.07	2.96	18	60	22	24
	3	7.33	2.57	0.5	66.5	33	24
	3	5.73	2.76	45	34	21	25
IV							
Rhythmically							
bedded		7.43	2.62	1.3	58.7	40	22
Silt and clay							

shorter time scales, the poor to very poor sorting of the cross-laminated beds indicates the waning of each current and the deposition of progressively finer sediment (Table 5.2).

The final phase is reached when underflow velocities decrease below the threshold for bed traction (< 20 cm/sec), and silt-clay laminations are deposited. At the Christie section, silts and clays settled-out to form rhythmic, as opposed to random laminations, at least 1 km from the meltwater input (e.g. section 22, Plate 4.15). Evidence of changing velocities can also be identified within the silt-clay rhythmites. Thin, fine sand horizons deposited between the rhythmites infer stronger currents, which in more proximal areas, were probably responsible for upper-stage plane beds within cross-laminated sediments. It is likely that both peak discharges, and mass movements near the lake margin, were directly responsible for the turbidity currents which deposited coarser bedforms amongst facies characteristic of lower flow velocites (Shaw, 1977; Shaw and Archer, 1978).

The silt-clay rhythmites grade upward into ripple cross-laminated, silty sands near the top of the Christie section (e.g. section 21, Figure 4.4). This is indicative of delta progradation, which locally decreased lake depth and produced a higher energy environment. At the top of the section there is an abrupt change from these cross-laminated silty sands to cross-stratified sandy gravels, which represent the final stage of deposition (e.g. section 17, Plate 4.12). This top 1 m unit was deposited in a high energy sandur environment as Plateau Lake lowered below the deposit surface. Meagre development of the glaciofluvial sequence capping the Christie section suggests that the lake drained quickly.

The stratigraphy and sedimentology indicate that this part of Plateau Lake was affected by more powerful meltwater currents than the southwest valley. This was due to a closer ice margin, however, the lack of ice-rafted sediments interbedded within the four facies, implies that the ice margin was not in contact with the lake. Alternatively, permanent lake ice may have prevented the migration of any icebergs from an ice-contact margin. It is likely that some dropstones deposited from the lake ice are scattered throughout the Christie section, however, none were observed within the individual sections.

The prominent transition of the underflows, with increasing distance from the meltwater input, is largely due to a deeper lake environment. This caused a loss of momentum and greater mixing of the underflows with the ambient lake water, which contrasts with the southwest valley. The poorer sorting of the Christie sediments (Table 5.2) indicates the pulsating nature of the underflows, however, they remained dominant throughout the existence of Plateau Lake.

Alternative modes of meltwater inflow, for example interflows and overflows, are not considered to have been significant in the sedimentation process. Due to its high arctic location, and colder conditions during the last glaciation (England and Bradley, 1978), Plateau Lake would likely have been permanently ice covered and therefore isothermal, remaining well below +4°C (maximum density of water). The meltwater was probably the same temperature or slightly warmer than the lake water, and due to its higher sediment content it would have had a higher density and therefore entered Plateau Lake as underflows. At the beginning and end of melt seasons, small discharges of colder water containing little sediment, could have had the same density as the lake. Due to its isothermal structure,however, meltwater of equal density would not have entered as interflows but as homopycnal flows (Bates, 1953). It is considered likely that the underflows were able to disperse sufficient fines into suspension to produce the silt-clay rhythmites predominantly by vertical settling during still conditions. Therefore, it is not necessary to invoke other types of inflow to explain the sedimentology at the Christie section.

Fine sediment derived from aeolian sources is not considered to be a significant component of the silt-clay rhythmites. The permanent ice covering Plateau Lake would have prevented an annual contribution of aeolian sediment, which is characteristic of temperate glacial lakes. The severity of sea ice in Clements Markham Inlet during the Holocene is indicated by the penetration of driftwood. No driftwood penetrated prior to 8,900 BP and the occurrence of driftwood is sparse between 8,900 BP and 6,000 BP as compared with the present (Stewart and England, 1983). More extensive sea ice implies a cooler climate during both the

last glaciation and the early Holocene. Therefore, assuming Plateau Lake existed during the last glaciation, the lower temperatures would have resulted in more extensive, possibly landfast lake ice, as compared with present High Arctic lakes. Furthermore, very little aeolian sediment was observed on South Lake, and no aeolian landforms were observed within the study area. This suggests that wind does not have an important role in sediment distribution in southern Piper Pass and supports the consideration that aeolian sediments are no more than a minor component of the Plateau Lake sediments. A section on the 'Chronology' of Plateau Lake follows later in this Chapter.

5.2.2 The Central Valley Section

The substantially smaller Central valley section is located approximately 2.5 km west of the Christie section (Figure 4.1). Its location, farther from the Divide Glacier, implies a more distal environment than the Christie section. The Central valley sediments, however, consist of fine sands, silts and clays. This indicates that low velocity underflows, comparable to the west end of the Christie section, were responsible for deposition at this section. Furthermore, ripple cross-laminated, silty sands composing the top of the section indicate that the paleocurrents flowed northward, down Central valley towards Piper Pass. Therefore, the sediments are considered to have originated from within the Central valley basin, rather than from the Christie section area of Plateau Lake.

The Central valley section is predominantly composed of planar-laminated sands and silts (Facies II). The section is not extensive enough to show spatial changes, however, the interbedding of the sands (subfacies 1) and silts (subfacies 2) demonstrates that the underflow velocities varied through time. The prevalent stratification is planar-laminated silt (subfacies 2), which indicates that the majority of underflows remained below 15 cm/sec and were not powerful enough to transport sediments as traction. In addition, the very poor sorting of the silts (Table 5.3) suggests that these low velocity underflows waned, depositing progressively finer sediment.

TABLE 5.3

Central valley section 10 Sections, 23 Samples

Facies	Subfacies	Graphic Mean ø	Inclusive Graphic Standard Deviationø (Sorting)	Percentages Sand	Silt	Clay	Section #
I	3	5.45	2.93	64	13	23	1
Ripple	3	5.58	2.90	57	21	22	4
cross laminated	3	6.02	3.01	40	31	29	4
silty sand	3	5.72	3.05	57	18	25	5
II	1	3.57	1.13	78	22		2
Planar	1	3.28	1.16	80	20		3
laminated	1	3.4	1.59	74	26		5
sands and	1	3.45	1.22	78	22		8
silts	1	3.73	1.25	74	26		8
	1	3.07	0.8	87	13		11
	2	6.85	3.05	25	38	37	1
	2	6.28	3.27	41	22	37	2
	2	6.8	3.19	35	26	39	2
	2	6.4	3.07	35	31	34	3
	2	5.65	3.03	55	19	26	4
	2	6.83	2.83	8	57	35	7
	2	7.38	2.83	3	58	39	7
	2	6.08	2.86	39	30	31	8
	2	7.12	2.87	10	52	38	9
	2	7.2	2.89	10.5	50.5	39	9
	2	7.17	2.98	18	41	41	10
	2	7.9	2.68	1	52.5	46.5	10
	2	7.13	2.81	12	46	42	11

The sand horizons (subfacies 1) are also fine enough to have been deposited entirely from suspension by slightly faster underflows (<20 cm/sec). However, an alternative interpretation is that they are upper-stage plane beds deposited entirely from traction. The evidence in favour of the second interpretation is: 1) the sand horizons reach thicknesses of 8 cm and appear uniform throughout, and 2) the sands are significantly better sorted than the silts (Table 5.3). If the fine sands are upper-stage plane beds, then underflows between 80 and 120 cm/sec occasionally interrupted the relatively calm environment. The evidence opposing this interpretation is: 1) There is no ripple cross-lamination either side of the fine sand beds indicative of increasing and waning underflow velocity. Therefore, these powerful underflows must have been abrupt, well defined pulses, which passed through the Central valley site. 2) Under such conditions some scouring of the underlying silts, or the presence of rip-up clasts might be expected; however, none of these structures were observed at the section. The author considers the planar-laminated fine sands (less than 1 cm) as products of low velocity underflows, whereas the fine sand beds (several centimetres) are most likely upper-stage plane beds deposited by occasional high velocity turbidity currents.

Ripple cross-laminated sediments only occur near the top of the Central valley section (e.g. section 6, Plate 4.18), which suggests that for most of the time a deep water environment existed at this site. Sediment accumulation created a progressively shallower, and therefore higher energy environment, through which underflows were able to maintain sufficient velocity to transport some of the sediments in traction. Unlike the Christie section, the Central valley section is not capped by a thin, coarse, glaciofluvial sequence. Although, this may be a reflection of rapid drainage (to an elevation below the deposit surface), it is more likely a reflection of the inability of the Central valley stream to transport sediments larger than medium sand.

Upstream of the Central valley section there are no mounds of coarse sediment indicative of a former ice margin. In addition, no sand and gravel are apparent which might represent a former ice-proximal lake environment associated with the Central valley sediments.

The only evidence of ice-rafting is provided by occasional fine gravel clasts, which were probably deposited from lake ice rather than melting icebergs. In conclusion, the Central valley sediments are not considered to be ice-contact, and furthermore, meltwater discharge into this part of Plateau Lake was, on the whole, considerably less than that entering Plateau Lake east of the Christie section where the Divide Glacier is known to have terminated.

5.2.3 Overall Relationships

The evidence from within Piper Pass does not support the model of a greatly expanded South Lake extending onto the Hazen Plateau. Certainly, the recent changes in lake elevation (~28 m) were confined within southern Piper Pass, and South Lake reached no further than 3 km up the southeast and southwest valleys from the present lake margin. In addition, there is no evidence of shorelines or perched deltas significantly higher than the uppermost shoreline (28 m). Therefore, the glaciolacustrine sediments on the Hazen Plateau were deposited into a separate proglacial lake (Plateau Lake), which was dammed by ice that occupied the southeast valley.

Reconnaissance along Turnabout valley did not reveal any lacustrine deposits indicative of ponded meltwater. However, subdued glacial deposits were identified, including eskers and drumlinoids, which demonstrate that at one time ice from Turnabout Glacier flowed down the valley towards Turnabout junction (the confluence of Divide and Turnabout valleys, Figure 4.1). The most prominent glacial deposit is a large moraine, which extends almost right across the mouth of Turnabout valley (Figure 4.1). At its highest point the moraine is approximately 23 m above the southeast valley and 83 m higher than South Lake (Plate 5.4). Therefore it is concluded that Turnabout Glacier advanced into the southeast valley, blocking the drainage into southern Piper Pass and creating Plateau Lake. The extent of the moraine suggests that at the onset of deglaciation the ice stagnated within the confined valley mouth. Large amounts of till were slowly lowered to the valley floor producing the hummocky topography.

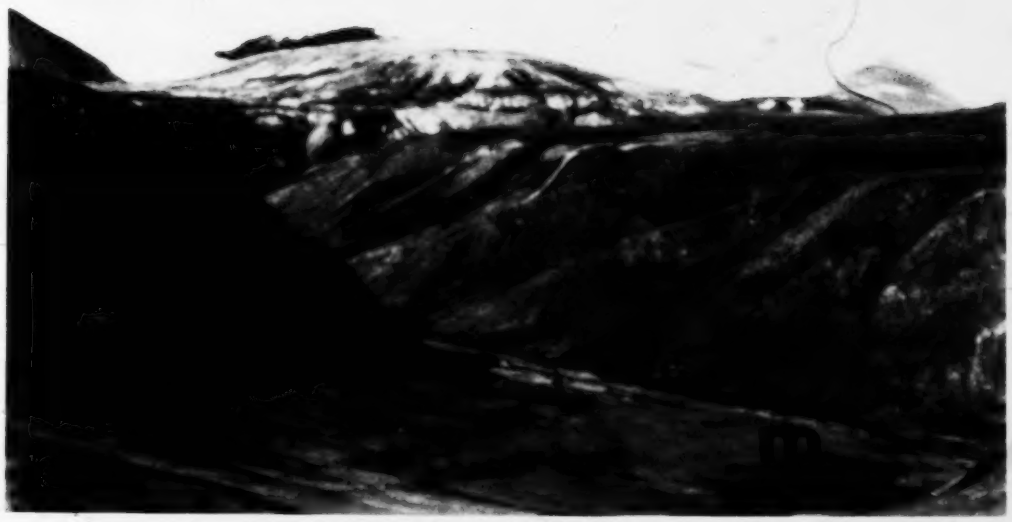

Plate 5.4 Looking down onto the Turnabout junction moraine (m), with the river occupying southeast valley flowing northwards (left) into South Lake. View taken from the top of a ridge above Turnabout valley.

A minimum estimate on the upper level of Plateau Lake is provided by the two sections whereas a maximum estimate is provided by related terraces. The main surface of the Christie section is approximately 156 m above South Lake. Two parallel terraces on the north side of Divide river, opposite the Christie section, are believed to represent former lake levels (Plate 5.3). The higher of the two terraces is 169 m above South Lake and is considered to represent the highest elevation of Plateau Lake. The second terrace is 10 m lower, and likely reflects an intermediate stage during the overall drainage of the lake. The top of the Central valley section is 145 m, which is well within the proposed 169 m maximum elevation for Plateau Lake. The lake could have attained this elevation (169 m) only if the ice damming Turnabout junction was at least 110 m thick. It is possible that the Turnabout Glacier advanced beyond Turnabout junction, along the Divide and southeast valleys; however, there are no ice-contact deposits on the valley sides which would support this.

The upper sediments of the Christie and Central valley sections indicate that Plateau Lake drained quickly to some elevation below their surfaces. However, the lower, western extension of the Christie section, along with lower terraces, demonstrate that the lake drained in several stages. This implies that ice occupied Turnabout junction, and fluctuated to adjust the size of the drainage outlet. The western extension is 143 m above South Lake, and signifies continued delta progradation during initial incision of both the Christie and Central valley sections (Plate 5.5).

Further evidence of this 143 m lake stage is provided by large sandstone blocks perched on this western extension of the Christie section (Plate 5.5). The blocks are not embedded in the sediment, therefore they are considered to have been ice-rafted by thick lake ice. The sandstone blocks must have originated from the outcrop of Permian sandstone, west of the southeast valley (Figure 1.3). Therefore, they were transported initially by Turnabout Glacier and then rafted by the lake ice to the Christie section. The capacity of lake ice to raft 2 to 3 m^3 blocks is clearly indicated on the southeast side of South Lake, where two equivalent blocks rest on the 28 and 16 m shorelines.

Plate 5.5 Looking east at the western extension of the Christie section which has sandstone blocks resting on the deposit surface. These are interpreted as ice rafted blocks in the former lake.

Downstream of the Christie section, on the north side of Divide river, additional minor terraces appear to have been formed by lower lake levels. These terraces, however, occur as four short, separate segments, which decrease in elevation from 113 to 102 m in the downstream direction. The indistinct nature of these terraces and their absence in other areas, warrants the alternative interpretation that they were formed by solifluction on the 20° valley side, rather than being lake shorelines.

The elevations of both the glaciolacustrine deposits and of related features, can be used to define the extent of Plateau Lake at its highest stage. In turn, this former lake and its sediments can be used to delimit the associated maximum ice margins during its existence (Figure 5.3). During the deposition of the Christie and Central valley materials the lake must have been linked, therefore Plateau Lake extended 4 km along Divide valley varying from 0.5 to 1.3 km in width . Divide Glacier likely terminated immediately east of the Christie section at some time during the lake's existence. This is suggested by the relatively high energy environment depicted by the proximal sediments and small moraines 0.6 km to the east of them.

5.3 Chronology

The ^{14}C dates from the ice-contact delta and southwest valley indicate that the recognizable fluctuations of South Lake occurred during the late Holocene. Previous glacio-climate studies indicate that the advance of the glacier causing the rise of South Lake to +28 m, may have been in response to the Little Ice Age (ca. 100 to 400 BP, Stewart and England, 1983) However, this is a minimum age because there could be a considerable time lag between the climatic change responsible for the ice advance (possibly much older than the Little Ice Age) and the actual advance causing the rise to 28 m. Furthermore, the retreat of the glacier from the lake's maximum stand, is dated by organics at 1585 ± 190 BP (although the sample is a <u>maximum</u> date for the delta). The aerial photographs indicate that South Lake has lowered 13 m since 1958. This recent lowering of the lake may have been caused by the

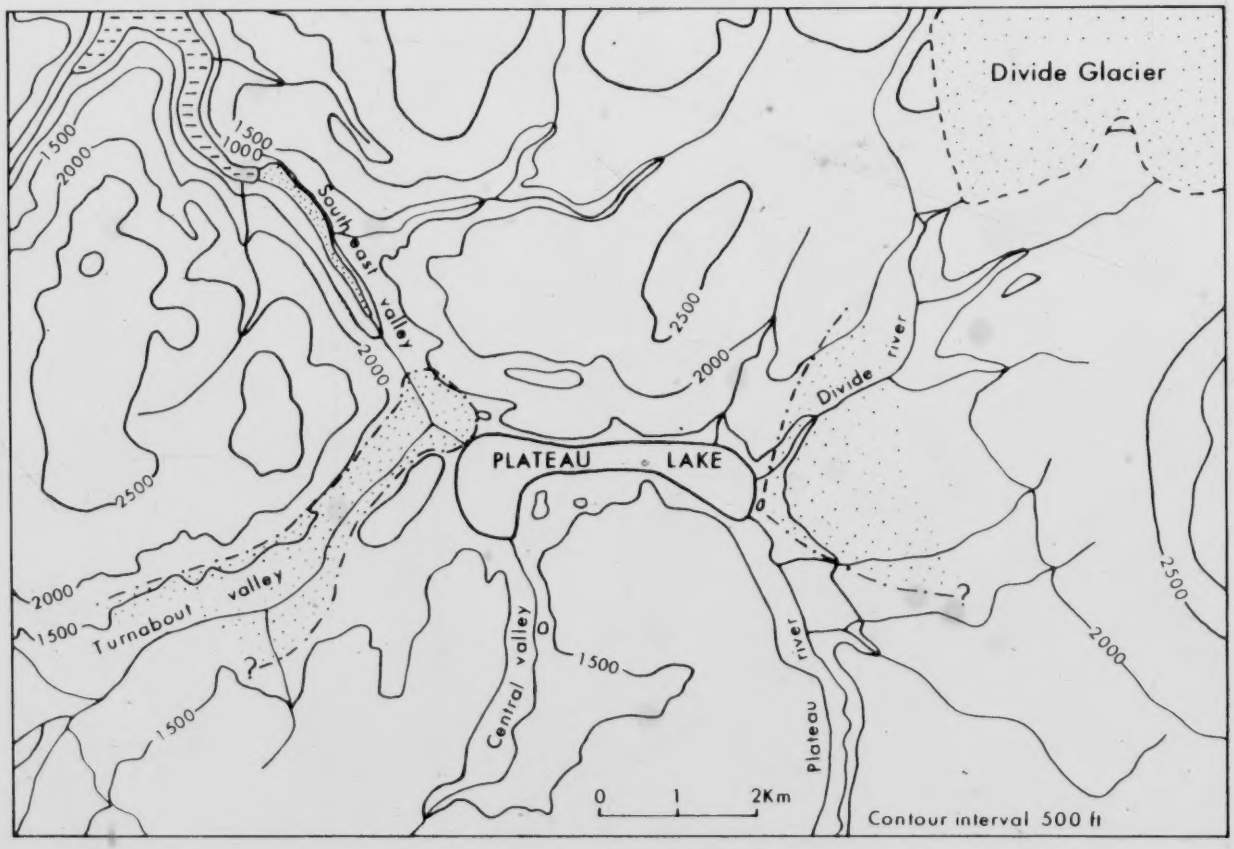

Figure 5.3 The approximate maximum positions of Turnabout and Divide glaciers during the existence of Plateau Lake.

prominent warming during the first half of this century that led to the retreat of the damming glacier (Hattersley-Smith and Serson, 1970; Fisher and Koerner, 1980).

The allochthonous organic material collected from the Christie and Central valley sections also provide maximum dates for Plateau Lake. The top sample from the Christie section (section 17) was dated at >33,000 BP.(S-2633), whereas the basal sample (section 16) was dated at 15,040 ± 625 BP (S-2632) Because these dates are stratigraphically out of order they suggest that the top sample is composed of a blend of organic material, possibly including wood fragments of Tertiary age. Tertiary wood is known to occur in this area and small fragments of redeposited *Salix* twigs in this upper sample may be of Tertiary age. For example, organics within a marine limit delta in St. Patrick Bay 70 km to the southeast (Figure 1.1), have been dated at 28,100 ± 380 BP (England, 1978). England (1978) noted that rounded wood fragments, possibly originating from Tertiary deposits around Lake Hazen, and an opaque yellow resin from the latter area, had been transported into the St. Patrick Bay deposits. Hence it is possible that the sample was contaminated although only the *Salix* sp. was suspected as being foreign. On the other hand, the basal, finite date from the Chrisite section indicates that Plateau Lake is possibly no older than 15,000 BP, although considerable sediment accumulation would have already occurred before the deposition at section 16. It is not known, however, to what degree this basal date reflects a blend of organic material of different ages. A small quantity of wood fragments was observed in the sample, therefore, if these are 'old' organics, Plateau Lake could be considerably younger than 15,000 BP.

The two ^{14}C dates from the Central valley section are also of full glacial age (greater than 8,000 BP, England, 1983) Organic material from near the deposit surface (section 6, Plate 4.18) was dated at 19,055 ± 2930 BP (S-2635) and a basal sample from section 11 (Figure 4.5) was dated at 20,950 ± 1200 BP (S-2634). If the dates represent the time of deposition in the lake, then the mean age difference (1895 years) is an indication of the duration of Plateau Lake. However the standard errors on these dates clearly indicate that they overlap and, this is due principally to the small sample sizes. Therefore, from a laboratory point of view the mean

ages must be considered provisional.

As with the Christie section, the degree to which these samples from Plateau Lake are a blend of different ages is not known. Therefore, caution should also be exercised when interpreting these dates and at present they must be treated as maximum ages. However, it should also be noted that the basal date (20,950 ± 1200 BP) from the Central valley section is based on a sample that contained only bryophytes and hence there is no apparent contamination from Tertiary wood fragments (C. Lafarge-England, personal communication, 1985).

The ^{14}C dates obtained from the Christie and Central valley organics indicate two possible ages for Plateau Lake depending on the composition of the dated samples. One possibility is that the three samples which provided finite dates are composed of plant material which was growing immediately before deposition. In this case, the dates represent the time of deposition and indicate that Plateau Lake is greater than 8,000 BP, therefore a full glacial lake. The second possibility is that all four organic samples consisted of a blend of various ages. This means the dates are older than the lake, therefore Plateau Lake could be postglacial in age (less than 8,000 BP). These two interpretations are discussed more fully in the following sections.

5.3.1 Plateau Lake - full glacial age

The first possibility, that Plateau Lake is of full glacial age, is very interesting in regard to the extent of ice on northern Ellesmere Island during the last glaciation. During the existence of this lake, Turnabout Glacier must have been at least 16 km beyond its present margin to have blocked the drainage through southeast valley. However, Divide Glacier could not have been more than 5 km beyond its present margin during the last glaciation because it terminated at, or just to the east of, Plateau Lake's eastern shore. Also, one cannot invoke an earlier period of ice retreat (just prior to the basal 15,000 BP date) because there is no evidence of corresponding emergence (glacial unloading) in the adjacent full glacial sea occupying both Clements Markham Inlet and the south side of the Hazen Plateau (England, 1983; Bednarski,

1984). Therefore, the glacial deposits immediately east of the Christie section likely define the last limit of the Divide Glacier regardless of the ^{14}C dates in the sections (i.e this position is based largely on the stratigraphy of the sections). During sedimentation into Plateau Lake, the areas south of both the Christie and Central valley sections are considered to have been ice-free (Figure 5.3). The Plateau and Central valley streams transported large amounts of organic material from these areas and the organics presently remain well preserved in the glaciolacustrine sediments. A full glacial age for Plateau Lake would have an important bearing on the approximate limit of the last glaciation suggested by England (1983) as shown on Figure 1.1.

5.3.2 Plateau Lake - postglacial age

The second possibility of a postglacial lake implies that the lake occupies terrain previously covered by more extensive ice during the last glaciation. Therefore, Plateau Lake was formed during deglaciation and its existence must have been controlled by the time of ice retreat, particularly by that of Turnabout Glacier which is considered to have dammed the lake. Previous research on the last glaciation of the Hazen Plateau indicated that Plateau Lake formed within the last ice limit, suggesting that it would be postglacial in age; however this ice margin was solely speculative in this area (cf. England, 1983). In Clements Markham Inlet, rapid glacial retreat had begun by 8,000 BP (Bednarski, 1984), whereas on the Hazen Plateau rapid retreat did not begin until 6,200 BP (England, 1983). It is not known whether the timing of glacial retreat in the study area corresponds to either of the above areas. Nevertheless, Plateau Lake probably would have drained by at least 6,200 BP when a prominent amelioration was in progress (Stewart and England, 1983). However, it may have drained much earlier.

For Plateau Lake to be of postglacial age, Divide Glacier must have retreated up Divide valley from Turnabout junction, possibly having previously coalesced with Turnabout Glacier. Plateau Lake may have formed simultaneously been and in contact with the retreating Divide Glacier. Therefore, the glacial deposits east of the Christie section would represent a stillstand

in the overall retreat of Divide Glacier. However, the sediments at both the Central valley and the Christie sections do not support this model of lake development. There does not appear to be any till underlying the glaciolacustrine sediments and ice-rafted sediments were not observed. Ice-contact deposits were not observed at the Central valley section. Finally, at the Christie section, the stratigraphy and sedimentology are indicative of proximal sedimentation adjacent to a stationary ice margin, for which there are observed moraines.

To comply with the sedimentology data, a postglacial Plateau Lake must have formed after Divide Glacier had retreated from its glacial limit, to a position east of the Christie section, while the Turnabout Glacier remained blocking the drainage to Piper Pass. This infers very different response times for the Turnabout and Divide glaciers, which is not supported by previous research on the local pattern of deglaciation. Generally, on the Hazen Plateau, glaciers are considered to have remained at or close to their glacial limits for several thousands of years (cf. from 11,000 to 6,200 BP) prior to rapid ice retreat and glacial unloading after 6,200 BP (England, 1983). Therefore, a model of Divide Glacier actively retreating while the Turnabout Glacier remained stationary (damming the lake), is anomalous to the present knowledge of deglaciation on the Hazen Plateau.

A final observation that may be significant is the presence of a small gorge on the south side of the Christie section (Figure 4.4, Plate 5.6). The gorge is cut into sandstone bedrock (Mesozoic) and is presently occupied by Plateau river. From previous observations at the ice-contact delta, the gorge appears to be a larger example of a bedrock channel formed by a lateral meltwater stream. This suggests ice occupied the immediate area at some time before sedimentation into Plateau Lake, possibly before Turnabout Glacier had reached the point where it could block this drainage into Piper Pass. However, whether the formation of this small gorge is related to the formation of Plateau Lake is not known. Lastly, if Plateau Lake is of postglacial age then dates younger than 10,000 BP should be forthcoming from its sediments.

Plate 5.6 Looking southeast into the small gorge on the south side of the Christie section, through which Plateau river is flowing.

5.3.3 Summary

The greater than 8,000 BP dates from the Christie and Central valley sections, suggest that Plateau Lake is of full glacial age. Furthermore, the lack of younger dates solely on bryophythes from the base of the Central valley section is considered supportive of a full glacial age for the lake. Therefore, contamination of Holocene material by Tertiary wood fragments in the Central valley dates is considered unlikely. A final point is that a large amount of 'old' material is necessary to contaminate an organic sample of Holocene age, whereas it takes relatively little 'young' material to produce a younger ^{14}C date on otherwise older material (Olson and Broecker, 1968). In this regard, it should be noted that the old date (>33,000 BP) from the upper part of the Christie section did contain a large amount of woody fragments that could be of Tertiary age. Because the Central valley and Christie sections are considered to have been deposited in the maximum stand of Plateau Lake, it follows that the presumably uncontaminated dates on samples from the Central valley represent the best age estimates for this former lake.

The sedimentology of the Christie and Central valley sections does not support the formation of Plateau Lake simultaneously with retreat of Divide Glacier. In addition, the only glacial deposits observed in the study area that mark a former margin of Divide Glacier are located east of the Christie section. Collectively, these summary points suggest a full glacial age for Plateau Lake, however, additional ^{14}C dates and further analysis of the organic samples are required to test this hypothesis.

5.4 Conclusions and Recommendations

5.4.1 Conclusions

The three ice-dammed lakes occupying the southern half of Piper Pass display shorelines up to 30 m above present lake level. Organic material collected from the raised lacustrine sediments of South Lake were dated at 1585 ± 190 BP (S-2637) and 495 ± 170 BP

(S-2636). The date of 1585 BP was obtained on organics collected from the ice-contact delta that was deposited into the highest lake level (28 m) and, as with all allochthonous material, it must be regarded as a maximum age for the delta and uppermost shoreline. Collectively, these two ^{14}C dates indicate that these changes in lake elevations occurred within the late Holocene. In addition, the absence of observable isostatic tilting along the uppermost shoreline of South Lake reflects its late Holocene age. Therefore, the surveying results on this shoreline do not add to the present knowledge of glacial unloading in the area, and can not be compared to the profiles on marine shorelines to the north in Clements Markham Inlet (Bednarski, 1984, in press) and to the south in Archer Fiord/Lady Franklin Bay (England, 1983).

The elevation of South Lake is controlled by fluctuations of the damming glacier at its north end. Lake expansion is considered to have been caused by the advance of this glacier and the closing of the lake outlet. This may have been in response to the substantial cooling of the Little Ice Age (ca. 100 to 400 BP). However, it should be noted that many glaciers on northern Ellesmere Island reached their maximum postglacial positions in the late Holocene in response to an earlier cooling after ca. 4500 BP (Stewart and England, 1983). Therefore the expansion of the glacier that dammed South Lake to +28 m may have resulted from a climatic deterioration older than the Little Ice Age. The very recent lowering of South Lake due to further opening of the drainage outlet, is thought to have been in response to the prominent warming during the first half of this century. The recently expanded South Lake did not extend onto the adjacent Hazen Plateau. Furthermore, there is no evidence of higher lake levels to suggest that South Lake has ever extended out of Piper Pass. Therefore, the older lacustrine deposits with the Divide and Central valleys are not related to a lake within Piper Pass.

A separate lake basin (Plateau Lake) formerly occupied the Divide and Central valleys. Plateau Lake developed when the northward drainage through southeast valley was blocked by the Turnabout Glacier. This ice dam remained at Turnabout junction long enough to form a proglacial lake nearly 4 km long and 100 m deep at its greatest extent. Plateau Lake varied in width from about 0.5 km in the confines of Divide valley, to 1.3 km at its west end where it

extended south, up the Central valley. It drained in several stages, finally leading to the erosion of lacustrine sediments deposited at the former ice dam maintained by Turnabout glacier.

Four maximum ^{14}C dates (> 33,000 BP, 15,040 ± 625 BP, 20,950 ± 1200 BP and 19,055 ± 2930 BP) obtained on terrestrial organics collected from the Christie and Central valley sections indicate that Plateau Lake could be of full glacial age. In addition, these ^{14}C dates suggest that Plateau Lake existed for at least 5,000 years. Therefore, Turnabout Glacier must have maintained its position at Turnabout junction over this length of time. Also, Divide Glacier could not have been more than 5 km beyond its present margin during the last glaciation (8,000 to 11,000 BP, England, 1983). A full glacial age for Plateau Lake would necessitate a significant change to the approximate limit of the last glaciation suggested by England (1983) as shown on Figure 1.1. A large ice free area would have existed on the Hazen Plateau immediately south of Piper Pass.

5.4.2 Further Research

This study has contributed to understanding the Late Quaternary environment of northern Ellesmere Island. The elevation changes of South Lake during the late Holocene exemplify a glaciological response to the recent, prominent climatic fluctuations identified in High Arctic ice cores (Koerner,1977; Paterson et al., 1977; Fisher and Koerner, 1980). The ^{14}C dates from the Plateau Lake sediments indicate that ice extent during the last glaciation may have been more restricted than previously thought. However, further research, both within the study area and along the southern Grant Land Mountains is recommended to refine the observations outlined in this thesis. These points are outlined below.

5.4.2.1 The Study Area

The most important aspect of further research is to resolve the accuracy of the existing ^{14}C dates regarding the absolute age of Plateau Lake (postglacial or full glacial). Given the possibility that these dates are mean estimates on samples containing a wide range of ages, it is necessary to test these "conventional" ^{14}C dates with accelerator dates

on individual fragments. This would provide two important points: 1) it would determine what the actual range of ages were on the material in the samples; and 2) the youngest ages would clearly provide the best, maximum ages on the deposition of these sediments. To this end, three samples have been sent to Isotrace Inc. (University of Toronto) for accelerator dates. Prior to this a botanist (C. Lafarge-England) separated what she considered to be the oldest (woody) fragments and the youngest (bryophyte) fragments from samples from the base and top of the Central valley section, which are stratigraphically equivalent to the samples previously dated. The youngest fragments from both the base and top of the Central valley section were submitted to Isotrace Inc., along with the oldest material (possibly Tertiary) from the top of the section. These dates are expected to be run in September, 1985.

5.4.2.2 The Lake Hazen Basin

On a regional scale during the last glaciation, trunk glaciers advanced southwards out of the Grant Land Mountains and into the Lake Hazen basin. The southward extent of these glaciers, and whether Lake Hazen was entirely or only partially occupied by ice remains unknown outside of the Ruggles River and Craig Lake areas (cf. England, 1983). Other glacial deposits are apparent on the aerial photographs and deglacial shorelines at ca. 100 m above the present lake have been noted by England et al., (1981). Given the abundance of organics in the Plateau Lake sediments, are there also organics in the Lake Hazen sediments? Careful mapping of ice margins and examination of the lake sediments should reveal 'The Lake Hazen Story'! Therefore, the Late Quaternary evolution of the Lake Hazen basin is proposed as an important, major extension to the research presented in this thesis.

BIBLIOGRAPHY

American Society for Testing and Materials (ASTM), 1982: *Soil Testing Procedures*, D-422: 114-120.

Antevs, E., 1925: Retreat of the last ice sheet in eastern Canada. *Geological Survey of Canada Memoir*, 146: 142 pp.

Ashley, G.M., 1975: Rhythmic sedimentation in Glacial Lake Hitchock, Massachusetts-Connecticut. In Jopling, A.V., and McDonald, B.C. (ed.), *Glaciofluvial and Glaciolacustrine Sedimentation*. S.E.P.M. Special Publication, 23: 304-320.

Banerjee, I., 1974: Sedimentology of Pleistocene glacial varves in Ontario, Canada. *Geological Survey of Canada Bulletin*, 226: 44 pp.

Bates, C.C., 1953: Rational theory of delta formation. *American Association of Petroleum Geologists Bulletin*, 37: 2119-2162.

Bednarski, J., 1984: Glacier fluctuations and sea level history of Clements Markham Inlet, Northern Ellesmere Island. Unpublished Ph.D. Thesis, University of Alberta, Edmonton, Alberta: 232 pp.

Bednarski, J., in press: The Late Quaternary in Clements Markham Inlet, Northern Ellesmere Island. *Canadian Journal of Earth Sciences*..

Bouma, A.H., 1962: *Sedimentology of some flysch deposits: A graphic approach to facies interpretation*. Elsevier, Amsterdam. 168 pp.

Brassard, G.R., 1971: The mosses of northern Ellesmere Island, Arctic Canada. Ecology and Phytogeography, with an analysis for the Queen Elizabeth Islands. *The Bryologist*, 74: 234-281.

Christie, R.L., 1967: Reconnaissance of the surficial geology of northeastern Ellesmere Island, Arctic Archipelago. *Geological Survey of Canada Bulletin*, 138: 50 pp.

Christie, R.L., 1976: Eocene beds at Lake Hazen, northern Ellesmere Island. *Geological Survey of Canada Paper*, 76-1C: 153-156.

Church, M. and Gilbert, R., 1975: Proglacial fluvial and lacustrine environments. In Jopling, A.V., and McDonald, B.C. (ed.), *Glaciofluvial and Glaciolacustrine Sedimentation*. S.E.P.M. Special Publication, 23: 22-100.

Clemmensen, L.B. and Houmark-Nielsen, M., 1981: Sedimentary features of a Weichselian glaciolacustrine delta. *Boreas*, 10: 229-245.

Coakley, J.P. and Rust, B.R. 1968: Sedimentation in an Arctic Lake. *Journal of Sedimentary Petrology*, 38: 1290-1300.

Crary, A.P., 1956: Geophysical studies along northern Ellesmere Island. *Arctic*, 9 (3): 155-165.

Crary, A.P., 1958: Arctic ice island and ice shelf studies, Pt. 1. *Arctic*, 11: 3-42.

Crary, A.P., 1960: Arctic ice island and ice shelf studies, Pt. II. *Arctic*, 13: 21-50.

Dansgaard, W., Johnson, S.J., Clausen, H.B., and Langway, C.C., Jr., 1971: Climatic record revealed by the Camp Century ice core. In Turekian, K.K. (ed.), *The Late Cenozoic Glacial Ages*. New Haven, Connecticut: Yale University Press, 37-56.

Dawson, A.G., 1983: Glacier-dammed lake investigations in the Hullet lake area, South Greenland. *Meddelelser om Gronland, Geoscience*, 11: 22 pp.

Day, P.R., 1965: Particle fractionation and particle-size analysis. In Black, C.A. (ed.), *Methods of Soil Analysis*. Agronomy, No. 9, Part 1. American Society of Agronomy, Madison, Wisconsin, 545-567.

Deane, R.E., 1958: Pleistocene geology and limnology. In *Operation Hazen:* narrative and preliminary reports for the 1957 season. Defence Research Board, Canada, 19-23.

Deane, R.E., 1959: Pleistocene geology and limnology. In *Operation Hazen*: narrative and preliminary reports, 1957-58. Defence Research Board, Canada, 61-63.

DeGeer, G., 1912: A geochronology of the last 12,000 years, International Geological Congress, XI Stockholm, 1910. *Comte Rendu*, 1: 241-258.

Domack, E.W., 1982: Sedimentology of glacial and glacial marine deposits on the George V - Adelie continental shelf, East Antarctica. *Boreas*, 11: 79-97.

Domack, E.W., 1983: Facies of late Pleistocene glacial-marine sediments on Whidbey Island, Washington: An isostatic glacial marine sequence. In Molnia, B. (ed.), *Glacial-marine sedimentation*. New York: Plenum Press, 535-570.

Drewry, D.J. and Cooper, A.P.R., 1981: Processes and models of Antarctic glaciomarine sedimentation. *Annals of Glaciology*, 2: 117-122.

Emerson, B.K., 1898: Geology of old Hampshire County, Massachusetts. *United States Geological Survey Monograph*, 29: 790 pp.

England, J., 1974: The glacial geology of Archer Fiord/Lady Franklin Bay northeastern Ellesmere Island, N.W.T., Canada. Unpublished Ph.D. Thesis, University of Colorado, Boulder, Colorado. 234 pp.

England, J., 1976a: Postglacial isobases and uplift curves from the Canadian and Greenland high arctic. *Arctic and Alpine Research*, 8: 61-78.

England, J., 1976b: Late Quaternary glaciation of the eastern Queen Elizabeth Islands, N.W.T., Canada: Alternative Models. *Quaternary Research*, 6: 185-202.

England, J., 1978: The glacial geology of northeastern Ellesmere Island, Northwest Territories, Canada. *Canadian Journal of Earth Sciences*, 15: 603-617.

England, J., 1982: Postglacial emergence along northern Nares Strait. In Dawes, P.R. and Kerr, J.W. (ed.), *Nares Strait and the drift of Greenland: a conflict in plate tectonics*. Meddelelser om Gronland, Geoscience, 8: 65-75.

England, J., 1983: Isostatic adjustments in a full glacial sea. *Canadian Journal of Earth Sciences*, 20: 895-917.

England, J., in press. The late Quaternary history of Hall Land, Northwest Greenland. *Canadian Journal of Earth Sciences*.

England, J. and Bednarski, J., in press: Postglacial isobases from northern Ellesmere Island and Greenland: new data. *Canadian Journal of Earth Sciences*.

England, J. and Bradley, R.S., 1978: Past glacial activity in the Canadian high arctic. *Science*, 200: 265-270.

England, J., Bradley, R.S., and Miller, G.H., 1978: Former ice shelves in the Canadian high arctic. *Journal of Glaciology*, 20: 393-404.

England, J., Bradley, R.S., and Stuckenrath, R., 1981a: Multiple glaciations and marine transgressions, western Kennedy Channel, Northwest Territories, Canada. *Boreas*, 10: 71-89.

England, J., Kershaw, L., LaFarge-England, C., and Bednarski, J., 1981b: Northern Ellesmere Island: A natural resource inventory. Parks Canada Report. The University of Alberta, Department of Geography. 237 pp.

Fisher, D.A. and Koerner, R.M., 1980: Some aspects of climatic change in the High Arctic during the Holocene as deduced from ice cores. In Mahaney, W.C. (ed.), *Quaternary Paleoclimate*. Norwich, England: GeoAbstracts, 249-271.

Folk, R.L., 1974: *Petrology of Sedimentary Rocks*. Hemphill Publishing Company, Texas. 182 p.

Gilbert, G.K., 1885: The topographic features of lake shores. *United States Geological Survey*, Fifth Annual Report, 69-123.

Gilbert, R., 1982: Contemporary sedimentary environments on Baffin Island, N.W.T., Canada: glaciomarine processes in fiords of eastern Cumberland Peninsula. *Arctic and Alpine Research*, 14: 1-12.

Gilbert, R. and Shaw, J., 1981: Sedimentation in proglacial Sunwapta Lake, Alberta. *Canadian Journal of Earth Sciences*, 18: 81-93.

Gilbert, R. and Church, M., 1983: Contemporary sedimentary environments on Baffin Island, N.W.T., Canada: Reconnaissance of lakes on Cumberland Peninsula. *Arctic and Alpine Research*, 15 (3): 321-332.

Glen, J.W., 1954: The stability of ice-dammed lakes and other water-filled holes in glaciers. *Journal of Glaciology*, 2(15): 316-18.

Gustavson, T.C., 1975: Bathymetry and sediment distribution in proglacial Malaspina Lake, Alaska. *Journal of Sedimentary Petrology*, 45 (2): 450-461.

Gustavson, T.C., Ashley, G.M., and Boothroyd, J.C., 1975: Depositional sequences in glaciolacustrine deltas. In Jopling, A.V., and McDonald, B.C. (ed.), *Glaciofluvial and glaciolacustrine sedimentation*. S.E.P.M. Special Publication 23: 264-280.

Harms, J.C., Southard, J.B., and Walker, R.G., 1982: *Structures and Sequences in Clastic Rocks*. Society of Economic Paleontologists and Mineralogists, Short Course Number 9. 249 pp.

Hattersley-Smith, G., 1957: The rolls on the Ellesmere ice shelf. *Arctic*, 10(1): 32-44.

Hattersley-Smith, G., 1960a: Some remarks on Glaciers and Climate in Northern Ellesmere Island. *Geografiska Annaler*, 42: 45-48.

Hattersley-Smith, G., 1960b: The ice cover of northern Ellesmere Island. *Annals of the New York Academy of Science*, 95: 282-289.

Hattersley-Smith, G., 1961: Some glaciological studies in the Lake Hazen region of northern Ellesmere Island. In Raasch, G.O. (ed.), *Geology of the Arctic*, Vol. II. Toronto: University of Toronto Press, 791-808.

Hattersley-Smith, G., 1963: The Ward Hunt Ice Shelf: Recent changes of the ice front. *Journal of Glaciology*, 4: 415-424.

Hattersley-Smith, G., 1969a: Glacial features of Tanquary Fiord and adjoining areas of northern Ellesmere Island, N.W.T. *Journal of Glaciology*, 8: 23-50.

Hattersley-Smith, G., 1969b: Recent observations on the surging Otto Glacier, Ellesmere Island. *Canadian Journal of Earth Sciences*, 6: 883-889.

Hattersley-Smith, G. and Serson, H., 1970: Mass balance of the Ward Hunt ice rise and ice shelf: A 10-year record. *Journal of Glaciology*, 9: 247-252.

Hattersley-Smith, G., and members of the expedition, 1955: Northern Ellesmere Island, 1953 and 1954. *Arctic*, 8: 2-36.

Hattersley-Smith, G., Fuzesy, A., and Evans, S., 1969: Glacier depths in northern Ellesmere Island: Airborne radio echo sounding in 1966. Defence Research Board, Ottawa, Canada. DREO Technical Note, 69(6): 23 pp.

Hughes, T., Denton, G., and Grosswald, M.G., 1977: Was there a late-Wurm arctic ice sheet? *Nature*, 266: 596-602.

Hutchinson, G.E., 1957: *A treatise on limnology*. Volume 1, part 1 - geography and physics of lakes. New York: John Wiley and Sons Incorporated. 540 pp.

Johnston, W.A., 1922: Sedimentation in Lake Louise, Alberta. *American Journal of Science*, 20: 237-287.

Jopling, A.V. and Walker, R.G., 1968: Morphology and origin of ripple-drift cross-lamination, with examples from the Pleistocene of Massachusetts. *Journal of Sedimentary Petrology*, 38: 971-984.

Koenig, L.S., Greenaway, K.R., Dunbar, M., and Hattersley-Smith, G., 1952: Arctic Ice Islands. *Arctic*, 5(2): 67-103.

Koerner, R.M., 1977: Devon Island ice cap: Core stratigraphy and paleoclimate. *Science*, 196: 15-18.

Kuenen, P.H., 1951: Mechanics of varve formation and the action of turbidity currents. *Geologiska Foreningens: Stockholm Fordhandlingar*, 73: 69-84.

Lafarge-England, C., 1985: Personal communication. Department of Botany, University of Alberta, Edmonton, Alberta, Canada.

Larsen, H.V., 1961: Runoff studies from a glacier in southeast Greenland. *Folia Geographica Danica*, 9: 158-160.

Leech, R.E., 1966: The spiders (Araneida) of Hazen Camp 81°49'N, 71°18'W. *Quaestiones entomologicae*, 2: 153-212.

Lemmen, D.S., 1984: Sedimentation in a glacially-fed arctic lake: Tasikutaak Lake, Cumberland Peninsula, Baffin Island, N.W.T. Unpublished M.Sc. Thesis, Queen's University, Kingston, Ontario. 157 pp.

Lewis, D.W., 1984: *Practical Sedimentology*, Hutchinson Ross, Stroudsburg, Pennsylvania. 229 pp.

Liestol, O., 1956: Glacier dammed lakes in Norway. *Norsk geografisk Tidsskrift*, 15: 122-149.

Lyons, J.B., Ragle, R.H., and Tamburi, A.J., 1972: Growth and grounding of the Ellesmere Island Ice Rises. *Journal of Glaciology*, 11: 43-52.

Lyons, J.B. and Mielke, J.E., 1973: Holocene history of a portion of northernmost Ellesmere Island. *Arctic*, 26: 314-323.

Maag, H., 1969: Ice dammed lakes and marginal glacial drainage on Axel Heiberg Island, Canadian Arctic Archipelago. Axel Heiberg Island Research Reports, McGill University, Montreal. 147 pp.

Marshall, E.W., 1955: Structural and stratigraphic studies on the Northern Ellesmere Island Ice Shelf. *Arctic*, 8: 109-114.

Mayewski, P.A., Denton, G., and Hughes, T., 1981: Late Wisconsin ice sheets of North America. In Denton, G. and Hughes, T. (ed.), *The last great ice sheets*. New York, NY: John Wiley and Sons Incorporated, 67-178.

McKeague, J.A., 1978: Manual on soil sampling and methods of Analysis. *Canadian Society of Soil Science*. 212 pp.

Miall, A.D., 1978: Permian stratigraphy at Piper Pass, northern Ellesmere Island, District of Franklin. *Geological Survey of Canada Paper*, 78-1A: 541-544.

Miall, A.D., 1979: Tertiary fluvial sediments in the Lake Hazen intermontane basin, Ellesmere Island, Arctic Canada. *Geological Survey of Canada Paper*, 79(9), 25 pp.

Middleton, G.V. and Southard, J.B., 1984: *Mechanics of Sediment Movement*. Society of Economic Paleontologists and Mineralogists, Short Course Number 3. 401 pp.

Olson, E.A. and Broecker, W.S., 1968: Sample contamination and reliability of radiocarbon dates. *Transactions of the New York Academy of Science*, Series 2, 20: 593-604.

Orombelli, G. and Gnaccolini, M., 1978: Sedimentation in proglacial lakes: a Wurmian

example from the Italian Alps. *Zeitschrift fur Geomorphologie*, 22: 417-425.

Paterson, W.S.B., et al., 1977: An oxygen isotope climatic record from the Devon Island ice cap, Arctic Canada. *Nature*, 266: 508-511.

Petryk, A.A., 1969: Mesozoic and Tertiary stratigraphy at Lake Hazen, northern Ellesmere Island, District of Franklin. *Geological Survey of Canada Paper*, 68-17.

Pharo, C.H. and Carmack, E.C., 1979: Sedimentation processes in a short residence-time intermontane lake, Kamloops Lake, British Columbia. *Sedimentology*, 26: 523-541.

Postma, G., Roep, T.B., and Ruegg, G.H.J.,1983: Sandy-gravelly mass-flow deposits in an ice-marginal lake (Saalian, Leuvenumsche Beek Valley, Veluwe, The Netherlands), with emphasis on plug-flow deposits. *Sedimentary Geology*, 34: 59-82.

Powell, R.D., 1981: A model for sedimentation by tidewater glaciers. *Annals of Glaciology*, 2: 129-134.

Prest, V.K., 1952: Notes on the geology of parts of Ellesmere and Devon Islands, N.W.T. *Geological Survey of Canada Paper*, 52-32.

Shaw, J., 1977: Sedimentation in an alpine lake during deglaciation, Okanagan Valley, British Columbia, Canada. *Geografiska Annaler*, 59A: 221-240.

Shaw, J. and Archer, J., 1978: Winter turbidity current deposits in Late Pleistocene glaciolacustrine varves, Okanagan Valley, British Columbia. *Boreas*, 7: 123-130.

Shaw, J., Gilbert, R., and Archer, J., 1978: Proglacial lacustrine sedimentation during winter. *Arctic and Alpine Research*, 10: 689-699.

Smith, D.I., 1959: Geomorphology. In *Operation Hazen:* narrative and preliminary reports, 1957-1958. Defence Research Board, Canada: 58-60.

Smith, D.I., 1961: The glaciation of northern Ellesmere Island. *Folia Geographic Danica*, 9: 224-234.

Smith, N.D., Vendl, M.A., and Kennedy, S.K., 1982: Comparison of sedimentation regimes in four glacier-fed lakes of western Alberta. In *Research in glacial, glaciofluvial and glaciolacustrine systems*. 6th Guelph Symposium on Geomorphology. Geo Books, Norwich, 203-238.

Stein, R., 1985. Rapid grain-size analyses of clay and silt fraction by Sedigraph 5000D: Comparison with Coulter Counter and Atterberg methods. *Journal of Sedimentary Petrology*, 55 (4): 590-593.

Stewart, T.G., in preparation: Deglacial-marine sediments in a High Arctic Fiord. Unpublished Ph.D. Thesis, University of Alberta, Edmonton, Alberta.

Stewart, T.G. and England, J., 1983: Holocene sea ice variations and paleoenvironmental change, northernmost Ellesmere Island, N.W.T., Canada. *Arctic and Alpine Research*, 15: 63-87.

Sturm, M. and Matter, A., 1978: Turbidites and varves in Lake Brienz (Switzerland): deposition of clastic detritus by density currents. In Matter, A. and Tucker, T.E. (ed.),

Modern and Ancient lake sediments. International association of sedimentologists, Special Publication 2: 147-168.

Theakstone, W.H., 1976: Glacial lake sedimentation, Austerdalsisen, Norway. *Sedimentology*, 23: 671-688.

Thomas, G.S.P. and Connell, R.J., 1985: Iceberg drop, dump, and grounding structures from Pleistocene glacio-lacustrine sediments, Scotland. *Journal of Sedimentary Petrology*, 55 (2): 243-249.

Thorarinsson, S., 1939: The ice dammed lakes of Iceland with particular reference to their values as indicators of glacier oscillations. *Geografiska Annaler*, 21: 216-242.

Trettin, H.P., 1971: Geology of lower Paleozoic formations, Hazen Plateau and southern Grant Land Mountains, Ellesmere Island, Arctic Archipelago. *Geological Survey of Canada Bulletin*, 203: 134 pp.

Trettin, H.P. (*Coordinator*), 1972: The Innuvitian Province. In Price, R.A. and Douglas, R.J.W. (ed.), *Variations in tectonic styles in Canada*. Geological Association of Canada, Special Paper Number 11: 83-179.

Weidick, A., 1963: Ice margin features in the Julienehab district, South Greenland. *Gronlands geologiskė Undersogelse Bulletin* 33, 133 pp.

Wetzel, R.G., 1975: *Limnology*. W.B. Saunders Company. 743 pp.

Wright, R. and Anderson, J.B., 1982: The importance of sediment gravity flow to sediment transport and sorting in a glacial marine environment: Eastern Weddell Sea, Antarctica. *Geological Society of America Bulletin*, 93, 951-963.